THE STRUCTURAL AND METAMORPHIC HISTORY OF THE WILMOT AND FRANKLAND RANGES, SOUTH - WEST TASMANIA.

by

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Except as stated herein this thesis contains no material which has been accepted for the award of any other degree or diploma in any university, and, to the best of my knowledge and belief, this thesis contains no copy or paraphrase of material previously published or written by another person, except when due reference is made in the text of the thesis.

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"Those who wish for satisfactory foundations of facts on which to build their theories, must even be content to take their hammers in their hands, and having strapped on their knapsacks, to seek in the field of nature, the facts for themselves".

Haughton, 1856.

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ABSTRACT

Tidally dominated, shallow-shelf sea conditions are indicated for the deposition of much of the late Precambrian sequence of the Frankland and Wilmot Ranges, Southwest Tasmania. Predominantly pelitic units may reflect tidal-flat or deltaic situations, and detritus in marine quartz sand horizons was derived from an environment of prevailing aeolian conditions perhaps by transgression.

Overprinting proves five cleavage-forming deformation events followed by at least two conjugate kink sets and a regional 'fault-drag' rotation of all structural surfaces. The major geometrical features are the result of D_1 and D_4 whilst D_2 and D_3 locally produce macroscopic folds; all D_1 to D_5 events are essentially coaxial. Major D_4 folds are generally tight upright structures with wavelengths and amplitudes of about four kilometres and they rotate all earlier events. Gravitational gliding was responsible for the emplacement of F_1 structures which were subsequently, in the main, only slightly modified by further overriding during the same event. D_1 folds root to the west and southwest and though F_2 show the same direction of overthrusting it is uncertain whether D_1/D_2 is a continuous event. D_1/D_2 show orthotectonic characteristics whilst D_4 is of a paratectonic style. D_5 and the later conjugate kink bands are minor in scale. The regional rotation of deformations before and including those of D_5 and the kinks is considered to have been caused by pre-Ordovician transcurrent movement on the Lake Edgar Fault. Quartz arenite, in D_1, first deformed by plastic

deformation in hydrolytically weakened diagenetic quartz overgrowth but soon stress difference, in the grain-supported arenite, was taken up by intragranular plasticity of the detrital grains. Structural grains become apparent in outcrop at less than 10% shortening and good cleavages require less than 20% shortening. Penetrative fabrics developed in pelitic rocks in D_1. All phases are extremely heterogeneous, with unstrained zones surviving to the present day.

Post-D_1 cleavages usually involve microfolding of earlier fabrics, pressure dissolution and/or intracrystalline plasticity of quartz and mica. The relative importance of each mechanism is dependent on the pre-existing fabric and mica content of the rock.

Investigations of strain in quartz arenite revealed a need to measure sedimentary fabrics from non-orogenic areas to provide a sound basis for work in deformed material where initial marker ratios and orientations were variable. Methods chosen must also allow for an independent check on the validity of two-dimensional strain ratios. The geometry of deformed cross-bedding shows that flexural slip was important in the formation of major F_1 structures which were modified by an average 25% flattening.

As sedimentary structures are commonly modified or mimicked by deformation, care in interpretation is emphasised. Soft sedimentary, pre-tectonic clastic dykes are often planar and sub-parallel to cleavages where both structures are at an angle to bedding. Ready convergence of sedimentary and tectonic elements is thus demonstrated and the use of the approximate parallelism of dykes and cleavage to support tectonic dewatering is considered unsound.

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CHAPTER ONE

INTRODUCTION

1.1: Nature of the Study

The study was primarily a structural characterisation of the 'metamorphosed Precambrian' rocks of the Frankland and Wilmot Ranges, Southwest Tasmania (see Figure 1.1). Basic features such as the structural sequence and geometry, and metamorphic conditions were to be studied with follow up investigations on aspects requiring more detailed analysis. The region analysed was considered to be within the Tyennan Geanticline which is the belt of metamorphosed rocks extending from just north of Cradle Mountain (415385) to the south coast (see Figure 1.2). At the commencement of this work, the eastern margin of the Tyennan Geanticline was poorly defined and mapping between Frenchmans Cap (405320) and Port Davey (415205) was virtually non-existent. 'Metamorphosed Precambrian' rocks were believed (Hydro-Electric Comm., unpublished reports) to outcrop throughout the area prior to the present study - such rocks were considered to be mechanically layered and highly recrystallised, with very sparse preservation of sedimentary features. Detailed mapping proved common preservation of sedimentary features and a wide range of strain states in an area of polyphase deformation suggesting a blurring of the metamorphosed and 'comparatively unmetamorphosed' subdivision applied to Precambrian rocks in Tasmania (e.g. 1976 1:500,000 Geol. Map of Tasm. Dept. of Mines). Fine detail of sedimentary structures could be studied in outcrops containing iron-rich almandine garnets produced under high greenschist or lower amphibolite facies conditions. Occurrences of sedimentary features were sufficiently common to allow an interpretation of the depositional environment.

Analysis of the geometry of deformed sedimentary structures developed into a major component of the project and displayed the need for comparing completely untectonised regions with their deformed equivalents. This principle was particularly followed in the study of strain in partly recrystallised quartz arenites.

The Franklands/Wilmots study was decided upon in November 1970 but mapping did not get under way until April 1971 because of access problems. To satisfy the author's interests in penetrative deformation, the study area was to be restricted to the 'metamorphosed Precambrian' rocks and as such was to be the first Ph.D. project devoted largely to these rocks. The region between Lake Pedder (435245) (pre-1971) and Strathgordon (425265) proved inviting on several counts. The northern end, between the Gordon and Serpentine Dams, was the scene of a major engineering undertaking which could provide considerable information from road cuts, quarries, tunnels, shafts, dam foundations, and the like. A road also came close to the southern end but the prospect of eventual water access to the length of the Ranges was more encouraging. Of particular appeal was the glaciated, highland nature of the terrain which, despite logistic problems, was known to provide extremely good outcrop and the chance to establish clear structural correlations in an area known for structural complexity.

As this portion of the Tyennan Geanticline had recently been made accessible, it was considered a useful link in a chain of studies involving Cradle Mountain (Gee, et. al., 1970), Frenchmans Cap district (Spry, 1963 G; Gee, 1963; Turner, 1971) and the Port Davey area (Spry and Baker, 1965; Maclean and Bowen, 1971). Between Frenchmans Cap and Cradle Mountain, time relations between the peak of metamorphism and structural events vary and further trends might be defined to the south.

1.2: Location and Access

The area mapped covers the south western corner of the Tasmania Lands and Surveys Wedge Sheet (1:100,000) and, the central and southern, eastern margin of the Olga Sheet. Generally the main ridge of the Frankland and Wilmot Ranges was mapped and all major ridges that led down to the old Serpentine Valley. When weather permitted, traverses were also made of ridges to the west but such excursions were few. With more support, several complete cross-range traverses could have been made, though the structural interpretation was not unduly hampered by their absence, as the major hinge lines cross the main backbone of the Franklands at low angles to give an overall view of the structure.

A strip of country approximately 40 km long and averaging 5 km wide was investigated (Figure 1.3; future grid references apply to this figure which uses a kiloyard grid), though terrain and vegetation played a major role in deciding in detail what was to be mapped. In the first summer season of mapping (1971-72), helicopter support provided by the Hydro-Electric Commission gave invaluable assistance in the mapping of the central, rugged portion of the area. Without the helicopter, specimen collecting would have been severely curtailed as most camps were two days unladen walk from the nearest road. The enlarged Lake Pedder had just started to fill due to the completion of the Serpentine Dam (982350 Figure 1.3) and was at that time of no use for water-borne access.

The Gordon River Road, completed in 1967, runs through the northernmost tip of the region under consideration. Though this road to Hobart roughly parallels the trend of the Franklands and Wilmots it is on average 15 km away across Lake Pedder and a treacherous, button-grass swamp. The road to the Scotts Peak Dam initially provided foot access to the southern end but nearly a day's walk was needed to reach the edge of the area of interest. No settlements of any kind are found in the region but a small portion at the northern end has been the scene of intense activity during the construction of two dams by the Hydro-Electric Commission. The construction village of Strathgordon lies some 10 km to the east of the Gordon Dam (978400).

The greater part of the mapping was carried out at altitudes in excess of 800 m in an area of rugged topography and harsh climate. Cold, wet winters prevent effective field work which is, therefore, restricted to the November - March period except for work near the roads around the Gordon Dam. The summers are generally mild but rapid weather changes in mountainous terrain are dangerous, especially when base camps are up to two days walk from the nearest road. High country areas may be affected by mist, low cloud and rain for weeks on end even in summer. Total rainfall on the crest of the ridge is in excess of 250 cm. Camping periods of 2 - 3 weeks were spent on field work, but up to 75% of any one trip has been lost due to poor weather. The overall field time lost due to poor weather was not recorded, but it was of the order of 25% to 40%. Several areas were mapped from camps established by the side of Lake Pedder (which became navigable with care in late 1972), others were set up in the high country after boating to the nearest point on the Lake's edge. Most of the Wilmots to beyond Koruna Peak were mapped from ridge crest camps put in after walking from the Serpentine Dam site. Dense scrub along the eastern slopes of the Wilmots made water borne assaults on the crest of the Range difficult. The Frankland Range on its south-west and southern slopes is thickly clothed by scrub which prevented access to these areas. Extremely dense scrub is found in the glacial valleys running north-east from the central portion of the Frankland Range. Occasionally, ridges with thin scrub are found to run in a westerly direction. Southwest of the Companion Range is relatively clear, as is the west slope of the Wilmots but little time was available to traverse these areas.

When the project was initiated, and for some years later, only Imperial System scale maps were available. Most of the mapping was carried out on 1:15,840 or 1:7,920 scale maps though for the mapping around the dam sites 1" to 400' and 1" to 100' maps were used. Aerial photographs for most of the Ranges are poor in quality and give little assistance for mapping at the detailed scale used. Some more modern runs were completed as Lake Pedder filled but were of little help because of the general low plunge of the structures and restricted rock type range.

A total of about 35 weeks was spent in the field, and most of this time was on the Ranges proper. Some 5,000 structural readings were recorded from approximately 1,200 field stations. A little over 14,000 measurements were made in conjunction with the strain analysis program to quantify, length, width and angular relations of strain markers.

1.3: Geomorphology

Little time was devoted to the study of landforms or unconsolidated deposits of the region which lies within the 'fold structure province' described by Davies (1965). The physiographic features of the area were believed to be largely controlled by fold structures where valleys follow the strike of soft rocks and hard rocks form the ranges. The structural control is obvious but by Davies' figure 18, it was implied that the trends were Tabberabberan (Mid-Devonian) in age which is incorrect in much of the southwest. In the Franklands area the major upright folds that are responsible for the topographic trends are pre-Middle Cambrian. The major drainage system is the Gordon and Serpentine Rivers which are superimposed on the structural trends though the Serpentine shows much more concordance with the structure. On both sides of the Wilmot and Frankland Ranges, rivers run generally parallel to the strike of resistant rock types, but locally cross the trends and eventually join the important transgressive drainage channels. The very wide, flat floor of the Serpentine Valley must be quite an old feature and seems to accurately reflect the area of micaceous rocks for the most part. In some areas, however, the floor is transgressive over flaggy quartzite bands between Detached Peak (000298 Figure 1.3) and The Starfish (020247). River erosion must be responsible for the formation of the feature which has been modified by glacial activity. Prior to flooding, the Serpentine River was highly sinuous with oxbow lakes, suggesting a slow outlet erosion rate through the Serpentine Gorge (970365). Presumably the valley was being filled with detritus at a rate in excess of that which could be transported out of the Gorge.

Highland features of the region reflect the predominating influence of past glacial activity, though there is a very pronounced polarity to the Frankland Range in particular. Fluvial processes are very important in the southwest flanks of the Companion Range (050110) and the central Franklands. To the east, proof of past ice action is very much in evidence. Landforms typical of mountain glaciation are in abundance including cirques, arêtes, hanging valleys, steep sided fairly straight valleys without interlocking spurs and markedly stepped long profiles. Cirques near Double Peak (014163) have near vertical headwalls of over 400 m in height. The distribution and form of the major glacial features have been given by Derbyshire et.al. (1965) and Banks and Derbyshire (1970). From depositional landforms such as the arcuate end moraines opposite Double Peak, ice must have reached down to an altitude of 350 m and have been a little way from the break of slope into the valley floor. This is the limit proposed by Derbyshire et. al. (op. cit.) but on the basis of very large boulders (circa 30 tons), considered to be glacially derived, it is proposed that in an earlier peak of glaciation ice pushed at least a further 1½ km into the valley. The evidence for this is best seen in square 0319 where mounds up to 20 m in diameter occur in association with very large boulders which also probably underly the mounds. Other features suggest a greater ice coverage than indicated on the Glacial Map (Derbyshire, et. al., 1965), particularly the smooth ridge behind the glacial valleys between the Lion (057137) and Greycap (095106). Ice could well have flowed southwest over this ridge and have then been directed along the valley between the Frankland and Companion Ranges. Most of the smooth ridge is in phyllite which might account for its subdued shape but it certainly lacks the very jagged appearance of other ridges which appear to have suffered intense frost shattering above the snow-line. Detailed studies elsewhere in Tasmania record several pulses of glaciation which to some extent can be equated with cycles elsewhere in the Southern Hemisphere e.g. Chile (Bowden, 1975). The Franklands region appear to confirm the presence of marked fluctuations, though one episode of glaciation was thought responsible previously (Davies, J.L., in Spry and Banks 1962, p. 245).

Glacier flow was largely controlled by pre-existing valleys though enough ice was formed to over-ride some lower divides and possibly the main ridge itself. At the peak of glaciation, the system was to some extent a network with ice separating Tombstone Hill (050160) from the Franklands and the peaks of the Cupola (048148), The Lion, The Citadel (063129), and Murphys Bluff (074127) being Nunataks. Though zones of intense frost shattering have not been analysed in detail they could give more information on the extent of ice in the area. The generally smooth outline of the Tombstone Hill mass may indicate that it too was covered by ice.

Where the Franklands run east to west, glacial activity is concentrated on the southern side. Two very pronounced cirques are found here and Lake Surprise (144085) and the lakes below Frankland Peak (130088) are dammed by recessional moraines. Lateral moraines sweep down to the main valley from this area. Several hollows at the head of south flowing streams were also largely ice generated. On the north face of this portion of the Franklands the influence of ice action is less certain. The lower slopes are covered by a bouldery deposit where the boulders are very large (20+ tons) and the large bite into the Range north of Frankland and Secheron Peaks may have been a zone of ice accumulation. On most ridges, east-facing areas show more signs of ice action because on the wetter west slopes ablation was more effective. The Arthur Range trending west northwest has cirques and glacial valleys on both sides of the Range and the Franklands would be expected to be similar. Mt. Sprent (965326), on the Wilmots, has a near vertical cliff face on its eastern side in front of a fairly flat depression. It is proposed that this is a discrete cirque but depositional features have not been located.

After the important readvance, the presence of several arcuate recessional moraines at the outlet of each valley, records an irregular retreat with several still-stands. Bowden (1975) was able to demonstrate a later small readvance in high level areas of the West Coast of Tasmania but the nature of the present observations would not detect such an event. A prominent feature of the old Serpentine Valley was the gradual slope down from the Franklands to the northeast with the Serpentine River being on the northeastern margin of the valley. This relationship was brought about by glacial and fluvioglacial deposits being transported from southwest to northeast. The concentration of glacial activity from Coronation Peak (012180) to Greycap meant that this portion of the valley received more glacially derived detritus, which also led to the ponding of the old Lake Pedder.

Largely in post glacial times several very well developed alluvial fans have developed at the break of slope between the Franklands and the Serpentine Valley. Very good examples occur south of the old Lake Pedder and another is found between The Starfish and the Wilmots on the eastern side of this valley. Moraines have been cut down by rivers to drain virtually all parts of the glacial valleys. In many situations the break has occurred at the join between the lateral and end moraines of the main readvance.

1.4: Previous Literature

The Franklands/Wilmot region was inaccessible prior to the opening of the Gordon River Road. Climatic and access problems meant that most of the major peaks in the area were not climbed until the early 1950's and the two Ranges were not fully traversed until 1961. The above problems, together with a lack of potential for economic mineral deposits, has meant an absence of detailed geological investigation. During the period of early exploration in southwest Tasmania several traverses were made that impinged, or nearly so, on the present study area. Most of these made reference to the major rock groups of the area and speculated on their relationships (e.g. Johnston, 1888; Twelvetrees 1908a, 1908b, 1909), but are of little consequence in terms of the present investigation. Many passing references have been made to the Tyennan Geanticline, the major tectonic unit which encompasses the Frankland and Wilmot Ranges. Brief mention has been made of the ranges immediately north of the Wilmots by Scott (in Spry and Banks, 1962, p. 117), and Corbett (1970) comments on the topographic swing of the mountain chains from north/south to east/west.

Reconnaissance mapping was carried out by the Hydro-Electric

Commission in the vicinity of the proposed dam sites, from a helicopter supplied base, before the road was completed. Their mapping and logging program continued through to the present and has provided valuable basic data which would have otherwise been impossible to obtain because of the progress of engineering works. The only published work specifically on the present thesis area was by C. McA. Powell (1969a) which was based on three weeks mapping at the Gordon Dam site. For some years the Broken Hill Proprietary Company held exploration leases over the whole of southwest Tasmania and their initial project was a helicopter based mapping program of all the region (Hall, 1967). During this survey several short visits were made to different parts of the Frankland and Wilmot Ranges. The map that was subsequently produced was largely based on photo-interpretation by extrapolating from the scattered sample points.

Since this project was initiated, The Geological Survey of the Department of Mines has begun to map the Wedge Sheet. Though both the Huntley and Pedder Sheets (respectively, the north and south halves of the Wedge Sheet) were worked on at the start, the effort has now moved to completing the Huntley map. From 1971 to the present, two Honours projects, under the supervision of the author, have been carried out in adjacent regions (Williams, 1973; Duncan, 1976). In this same period the Hydro-Electric Commission geology section has produced several papers on the engineering geology of the dam sites (Roberts and Andric, 1975; Roberts, Cole and Barnett, 1975; Andric and Roberts and Tarvydas, 1976). Several publications of the author's relate specifically to the region now under consideration (Boulter, 1974a, b and 1977; Boulter and Raheim, 1974) and Boulter (1976) deals with studies that eventually permitted strain analysis of quartz arenite in the Franklands to proceed.

Several recent reviews deal with various aspects of Precambrian geology in Tasmania (Williams and Turner, 1974; Turner and Boulter, 1975; Williams, Solomon and Green, 1976; Williams, E., 1976). According to most writers the Tyennan Geanticline consists of rocks metamorphosed during the Frenchman Orogeny which is considered to be Upper Proterozoic in age (Raheim and Compston, in prep). The term 'metamorphosed Precambrian' has been applied to rocks of this tectonic element where they have been studied previously. The present trend is to use the terms 'metamorphosed' and 'unmetamorphosed' Precambrian in purely a descriptive sense, with the latter category being modified in the recent literature to 'comparatively unmetamorphosed'. Spry's statements on this type of subdivision have considerably influenced work in the Tasmanian Precambrian and he consistently favoured the interpretation of the metamorphosed Precambrian rocks being formed before the Frenchman Orogeny and the unmetamorphosed Precambrian being formed after the Frenchman (Spry in Spry and Banks, 1962, p. 123 and in Spry and Baker, 1965, p. 17). Though it was realised that such a stratigraphic implication was not proven (and still is not), the notion of equating metamorphosed with older and unmetamorphosed with younger Precambrian became established. The Penguin Orogeny is believed to be responsible for deformation in the 'comparatively unmetamorphosed Precambrian' of the northwest coast region and the orogeny with a minimum age of 720 m.y. (Richards, J.R. in Solomon and Griffiths, 1974) is considered to be younger than the Frenchman. According to present usage, the Arthur lineament, a high strain zone within the 'unmetamorphosed Precambrian', of the northwest, would be 'metamorphosed Precambrian'. Because of Spry's stratigraphic connotation for these terms confusion may arise but as yet Precambrian events in Tasmania are open to several interpretations through lack of contacts between the two distinct metamorphic groups and incomplete mapping.

It is realised that the intensity of the Frenchman Orogeny may vary from place to place thus producing an interleaving of the metamorphosed and unmetamorphosed Precambrian. Spry met this problem on several occasions (1963c, p. 11, 1964, and Spry and Baker, 1965) and invariably proposed a structural and metamorphic break between the two subdivisions. In the present study area delicate sedimentary structures and textures are found in close association with totally recrystallised quartz rich rocks. Strain analysis and textural investigation of quartzite in the Frankland area has shown strain and recrystallisation which elsewhere has been proven to be associated with quartz c-axis fabrics. On the basis of the fabric in many of the rocks, the area would be included in the 'metamorphosed Precambrian' but contains many features of the comparatively unmetamorphosed Precambrian. Wide variations in metamorphic and structural grade are, therefore, proposed for the Frenchman Orogeny. The question of the time relation between the Frenchman and Penguin Orogenies will only be resolved by careful radiometric age dating, closely related to detailed mapping.

1.5: Conventions

Throughout this thesis several standard formats have been established for the presentation of data. All structural readings are quoted with respect to true north and north marks on stereonets refer to this direction rather than magnetic north. All stereographic projections are equal area unless otherwise stated. When contoured, the diagrams show contours 95%, 50% and 25% of the maximum value found for each example.

Planar structures are recorded following the 'clockwise convention' where dip readings always appear as two digits and strike readings as three (e.g. 05.008). Of the two possible strike bearings, the one clockwise round from the dip direction is taken consistently.

The whole region is one of gentle plunges of minor hinge lines, which, as a result of refolding and initial variation of pitch within axial surfaces, tend to plunge in opposite directions within small areas. If one were to follow the normal practice of noting the asymmetry of the parasitic coupled folds (i.e. two folds with a short, common limb and two long, external limbs) as seen looking down the plunge of the fold, then a very complex pattern would result. Even for one generation of folds a map S, Z and M symmetrics would be difficult to interpret. With regular minor fold hinge-line plunges, an area between one major antiformal closure and its adjacent synformal closure would show a consistent S or Z pattern for minor structures of the same generation as the large structures. If congruent folds plunge in opposite senses, then in the same area S and Z would be found together, thus clouding the definition of the regular limb region. For this reason it was decided to indicate the asymmetry of minor folds as they appear looking north where the Ranges are north/south, north-west where they trend northwest/southeast and west where they are east/west. All field recordings for folds of every generation were changed to this standard form for representation on the map.

In the Elliott plots for strain analysis, a circle appears in the north-west quadrant of most plots to show the size of the counting circle used in the Mellis method. The digit inside the circle indicates the number of overlapping circles that were used to define the contour. The number or numbers beneath the circle indicates the diameter of the counting circle used and the size is quoted in units of the radius vector scale ($0.1 \equiv 0.1\varepsilon$). In a diagram presenting several sets of plots, if the above parameters are unchanged they will not be repeated for each plot. A two digit number in the south east quadrant indicates the angular relation between the measured section and bedding as a dihedral angle.

Grid references will be based on the kilo yard reference of the Imperial System maps. When reference is made to a specific locality from a detailed structural map area the grid reference will be in yards using the following format 41182 - 70973 with a dash separating easting and northing.

The following notation is used for structural elements: S_o bedding, D_n is the nth deformation event, F_n indicates folds produced during D_n, S_n a surface and L_n lineations produced during D_n. For descriptive purposes the following approximate size classifications have been used in this thesis to indicate amplitudes and wavelengths of folds: 1st order one kilometre +, 2nd order 100s metres, 3rd order 10s metres, 4th order metres. A minor fold occurs on an outcrop scale.

To describe individual folds the recommendations of Fleuty as modified by Ramsay (1967) have been followed.

All sections and profiles will be shown with the vertical and horizontal scales being equal unless, in exceptional cases, an exaggeration has been required. This latter condition will be clearly marked.

1.6: Acknowledgements

During a part-time study of this length one becomes indebted to many people and organisations for their various forms of assistance. Emyr Williams is to be thanked in particular for continual comments, criticisms, support, visits to the field and general encouragement through the whole study. Without his assistance the work would have been extremely difficult. Professor Carey, Dr. E. Williams and Mr. Gordon Hale arranged for my use of the Hydro-Electric Commission's helicopter in the 1971-1972 summer season which proved a tremendous boon to the investigation of the inaccessible central Franklands. Mr. Peter Williams, of the Geological Survey, whilst at the University for two years gave me the invaluable opportunity to test out many of my ideas on a critical mind. Discussions with Honours students working in the deformed rocks of Tasmania have also proved extremely helpful. Messrs. S.J. Williams, A.C. Duncan, S.F. Cox, D.B. Seymour, P.G. Lennox and N.J. Turner are in this category.

During my initial investigations in the Franklands region, Arne Raheim from the Australian National University became interested in the metamorphic petrology and Rb/Sr geochemistry of the Tasmanian Precambrian. His detailed chemical studies complemented my own structurally biased approach and led to fruitful cooperation. The knowledge gained from Dr. Raheim's study has been extremely useful in the evaluation of metamorphic conditions and a joint venture relating chemical variations to microtextural events was a logical outcome of this association (Boulter and Raheim, 1974). I am also grateful to the Australian National University for financially supporting a visit to Canberra during this project to discuss the matter with Dr. Raheim and to study the analytical techniques used.

For their ready interchange of information the geologists of both the Mines Department (N.J. Turner, A.V. Brown, M.P. and J. McClenaghan) and the Hydro-Electric Commission (G.J. Roberts, M. Andric, and R.K. Tarvydas), who have been working in the same region, are to be thanked. Both groups have contributed much to the evolution of my thoughts on the area.

At Strathgordon both Mr. and Mrs. R. Tarvydas and Mr. and Mrs. M. Dargaville extended their hospitality and the assistance of Murray Dargaville on Lake Pedder eased the burdens of camp life.

Over the time of the project I have taken several groups of geologists to the Gordon Dam area and their comments and discussion in the field have proved most valuable. Included in these visitors were Dr. D.G. Bishop, Dr. D.R. Gray, Dr. J. Griffiths, Mr. B.J. O'Connor, Dr. C. McA. Powell, Professor J. Rogers, Dr. M. Rubenach, Professor R.W.R. Rutland and Professor D.S. Wood.

Adrian Bowden kindly pointed out some of the glacial features of the district.

Members of the staff and graduate school of the University of Tasmania, Geology Department contributed much by way of discussion at seminars and on other occasions. In particular J. Walshe assisted with interpretations of metamorphic mineralogy. Andrew Duncan, Peter Cassidy and Victor Caune all acted as field assistants for periods of two months or more. I thank them for their good company in the bush and their strength in carrying equipment and specimens.

The Hydro-Electric Commission allowed access to the engineering workings at the Gordon and Serpentine Dam Sites and the National Parks and Wildlife Service allowed the work within the Southwest National Park. Many detailed base maps were supplied by the Hydro Electric Commission. Financial support came in part from the University of Tasmania Research Fund and the Department of Mines prepared some thin sections. The Photographic Section, University of Tasmania, is to be thanked for friendly assistance throughout the whole project.

My wife acted as a field assistant at times and is to be thanked for putting up with this project over several years and also permitting household funds to be diverted towards the project for field supplies, a boat, etc.

Professors S. W. Carey and D. H. Green provided facilities at the Geology Department, University of Tasmania. Professor D. H. Green very kindly arranged for much of the thesis to be typed at the University of Tasmania; that section was typed by Mrs. M. Martin.

CHAPTER TWO

ROCK TYPES

2.1: Metasediments - General Considerations

In this chapter rock names are used based on their inferred state either before or after lithification but prior to metamorphism. Structural complexity coupled with poorly outcropping micaceous rocks, renders a complete sedimentological analysis impossible. Large scale stratigraphical information concerning the relative abundance of different compositional types is not available and restricts a sedimentary interpretation of the metasediments. Another limiting factor is that near total recrystallisation has normally occurred in rocks that had initially more than about 2-5% argillaceous content. Sedimentary textures and mineralogies are, therefore, only well preserved in some quartz arenites. Flaggy siltstone, with minor mud content, tends to be completely recrystallised whilst medium-grained quartz sand with a muddy matrix may contain little-modified detrital grains. Interpretations of diagenetic histories, which might have provided critical information on the presence of supratidal suites, are rarely possible and always incomplete. Sedimentary structures are mostly obliterated in pelite, with only rare zones preserved showing sufficient information for interpretations of the local environment to be made. No explanation can be readily given for the local retention of fine structures in some pelite horizons and not in others, for the degree of strain, folding and metamorphism may be the same.

Within this section little attempt will be made to justify the identification of sedimentary, pre-lithification structures as opposed to tectono-sedimentary structures and purely tectonic structures mimicking features of undeformed sediments. A more complete discussion of these topics will be left to a later section (Chapter 6), dealing primarily with deformation of sedimentary structures. The tectono-sedimentary category is designed to include features that may be produced in a sequence being subjected to tectonic stresses before lithification e.g. clastic dykes and mud injections.

2.2: Metasedimentary Lithologies

As the discussion of the sedimentology is limited by the factors mentioned above, the description and analysis will be organised in terms of broad compositional groups rather than the standard facies devisions. Two basic lithological types are recognised, the predominantly quartz arenite type and the largely mudstone association. The rock distribution map (Figure 2.1) has employed a more detailed subdivision of lithologies.

2.2: i) The quartz arenite association. The bulk of rocks in this grouping are of pure quartz sand and, where mica is present, the original matrix content is rarely likely to have exceeded 10%. Coarse sandstone and pebbly conglomerate are quite common rock types between Murphys Bluff and Terminal Peak (44374, 46245, 46246, 46294). Some of the conglomerate layers appear graded (46257) but all are found in moderately to highly strained layers which limits recognition of structures. Rare conformable breccia layers occur with angular quartzite clasts up to 1½ x 3 cm in size set in a clean sand matrix (46258). Similar breccias occur in the Sentinels area, both as sedimentary layers and as discordant fault breccias (Williams, 1973).

The quartz arenite is texturally and mineralogically supermature (Folk, 1968). Measurements made in conjunction with the strain analysis program indicate that the initial range of axial ratios was often quite similar to that found in dune sand. The patterns (see Chapter 7) gave a very restricted range of ratios which is attributed to extensive abrasion in a wind dominated environment. Besides being of exceptional sphericity, the grains are very well sorted and extremely well rounded (Figure 2.2, a,b). Patro and Salu (1974) have shown that sorting is highest in the aeolian environment but contend that beach abrasion is most intense. The comparisons made in this study are based on the combination of sphericity and orientation measurements using the Elliott method (see Chapter 7). Totally unstrained specimens are rare, but it appears that, prior to deformation, the quartz grains were almost solely of the monocrystalline form (Figure 2.2 b). The absence of polycrystalline grains, which are more susceptible to breakdown in transport (Blatt and Christie, 1963), again indicates extensive abrasion. A high grade metamorphic source is also suggested by the exclusion of original polycrystalline grains. In the low to moderate strain examples, the original grain boundaries are often clearly delineated by dust trails of haematite. Grain shape parameters are, therefore, clearly seen along with the extensive development of quartz overgrowths on the detrital grains. The pure quartz sand would have possessed high initial porosity and the cement that now fills the pores must have come from dissolution of quartz grains either at high stress contacts (pressure solution) or by the general expulsion of pore fluids, in extremely quartz rich rocks. From time to time iron must have been carried in solution during diagenesis as several specimens record a complex filling of pore space (Figure 2.3). Some specimens show quite a high haematite content (Figure 2.4 a). Carbonate species were also present during diagenesis, the rare oolitic particles of specimen 46310 being interpreted as post depositional features (Figure 2.4 b). In the field, these carbonate micro-concretions are black in colour and are widely scattered through any one specimen.

Further evidence for extensive transport and reworking comes from the rounded grains of zircon and tourmaline (Figure 2.5). The tourmaline grains also show tourmaline overgrowths (Figure 2.5 a) similar to those described by Hemingway and Tamar-Agha (1975).

Herringbone cross-stratification is by far the most common sedimentary structure in the arenite beds of the region. The occurrence of cross-bedding is well distributed and the herring-bone patterns are almost always present, though one direction may predominate (Figure 2.6). All sets have sharp, planar boundaries of an erosional nature. Truncation surfaces are always inclined to give planar, wedge cross-bedding (Michelson and Dott, 1973) with sets normally persisting for less than 10 m. Trough cross-bedding is rare but this may be a function of the size of outcrop available. The set thickness is usually in the range 10-20 cm. The angle of inclination of the cross-stratification has not been used in the interpretation of the sedimentary environment because this parameter is readily modified by tectonic processes. Other sedimentary structures include washouts where the lower boundaries are curved surfaces of erosion (Figure 2.6 b). Water escapement structures are seen very rarely. Ripples are not common and, although they include asymmetric (directional) types, are so infrequently pre-served (or developed?) that they cannot be used rigorously in a palaeocurrent analysis. The ripples are slightly sinuous (Conybeare and Crook, 1968) and the lunate varieties have not been recorded. Several zones, some metres in thickness, of parallel laminated quartzite have also been noted.

Palaeocurrent analysis was made in areas where exposure enabled good control of the tectonic structures, such that zones of consistent facing and attitude were sampled. In this way the complications of parasitic folds were avoided and the analysis was limited to lowly strained uniform regions. The method of tectonic unwinding is discussed in Chapter 6. Apart from the three main sample sites, measurements were made at several other localities but lack of structural regularity often restricted measurement to about 10 per station. One of these stations was useful in showing that the herringbone pattern was related to modes 100° apart and was not necessarily bipolar (180° modes) bimodal, as might appear from many outcrops. On the whole the current patterns were so complex that the stations with few readings did not give meaningful results. The well-sampled sites were all from the same structural domain, thus allowing direct comparison of the bearings of the modes. Two samples show bimodal, bipolar patterns (Figure 2.7 a and c) with similar orientations though the second mode is very subdued in one case. The other sample site shows the same northwest and southeast modes but has two more to the northeast and southwest (Figure 2.7 b). The uncommon ripples in this area trend a little south of west suggesting that the northwest and southeast modes are the most important.

Examples of penecontemporaneous deformation of cross-bedding (Allen and Banks, 1972) have been recognised. These structures pose a considerable problem in this region for they may be morphologically identical to tectonic folds and, where cleavage parallels bedding, they may be indistinguishable. This difficulty is further discussed in the chapter dealing with deformation of sedimentary structures.

2.2: ii) The mudstone association. This association is characterised by a high proportion of mudstone but may at times be interlayered with up to 3 m thick sets of pure quartz sand layers. Regular interleaving of mudstone with quartz sandstone is uncommon and units are either of beds of sandstone with low matrix content or mudstone layers with subordinate sandstone. The mudstone association generally shows some preserved sedimentary structures, but localities where preservation is high are rare and interpretations are complicated by two phases of near isoclinal folding and one or two later events in most outcrops. A general list of sedimentary structures will not be given for this group because their variation and interaction, laterally and vertically, can rarely be determined. During the mapping program, however, many localities were found to show good preservation of delicate clastic dykes along with other sedimentary structures. Several of the best studied sites will now be described.

Locality 1 is on the Gordon River Road 76.8 km from Maydena (Figures 2.8 and 2.9). Part of the road section (between A and B on Figure 2.9) shows recognisable sedimentary structures and original layering of units of laminated grey to dark-grey argillaceous siltstone and black mudstone. Some of the coarser siltstone layers have ferroan carbonate contents of up to 40 percent (40636, 40641, 40684). In similar rocks, Anderton (1975) has been able to demonstrate a secondary replacement origin for carbonate which had pseudomorphed bladed gypsum crystals. The groups of these components range in thickness from 1 to 2 m down to a few millimetres, though they commonly are of the order of 2 to 8 cm (Figures 2.10 and 2.11). Parallel and cross-laminated sedimentation units are the most frequently observed sedimentary structures; scour and fill features (Figure 2.10) are dominant in the sequence. The steep sided scours are filled with parallel laminated and cross laminated silt (Figure 2.12 a). Lenticular bedding grading towards flaser bedding (Reineck and Wunderlich, 1968) is present in the section. Many units have a graded nature, and in thin section (40681, 40684) of finely laminated siltstone and mudstone, irregular contacts in the form of load casts and flame structures are developed. Rare examples of convolute folding, intraformational slump folds, and breccias have been noted.

A common structure is multiple, contorted siltstone sheets (Figure 2.11 and 2.13) most of which are between 2 and 6 cm long and 2 mm wide with some attaining a thickness of 1.5 cm. Their tabular form can be determined from linear intersections with bedding (Figure 2.14). These tabular structures have complex, branching forms generally at a high angle to the compositional layering, though occasionally planar offshoots run parallel to the layering. In this respect, they are similar to sandstone sheets described by Truswell (1972) from South Africa. In all cases, where facing evidence is available, the greater proportion of the clastic dykes intrude stratigraphically downward; only rare cases of upward intrusion have been determined (40645). Closely associated or continuous with the dyke and sill forms are bedding perpendicular zones of structureless siltstone (Figure 2.13 and 40628) in an otherwise finely laminated unit. Some clastic dykes originated from source beds that were subsequently eroded. This, together with some upward intrusion and examples of intersecting siltstone bodies, is taken to indicate a mode of formation that involves the forceful intrusion of silt in a liquified state into the adjacent mudstone, rather than particle by particle filling of a gaping fracture (compare with the form of structures described by Donovan and Foster, 1972). Some of the dyke/bedding intersection patterns may suggest modified sun-cracks or synaeresis cracks.
Of all the possible origins for clastic dykes (see Table 1 of Dionne and Shilts, 1974), it is believed the cause was shock, either by load due to sediment deposition or earthquake activity, producing spontaneous liquefaction of sand layers and fracturing of the adjacent

mud. Several phases of intrusion can be determined where dykes of slightly varying compositions cut one another (see point A, Figure 2.13). The contortions of the dykes suggest compaction after intrusion. Timing and conditions of formation will be further discussed in a later section.

Locality 2 is exposed in a generally pelitic zone outcropping in the saddle between Cleft Peak and Greycap (Figure 2.8). Sedimentary structures appear to have been partially obliterated by tectonic deformation, and although lenticular sedimentation units are common in an originally mudstone/siltstone sequence, little internal cross-lamination is preserved.

In restricted zones, thin (2 to 6 mm), multiple, planar dyke-like structures extend for up to 25 cm down from 0.3 m thick quartzite layers (Figures 2.15 and 2.16). Also, straight tabular dykes (averaging 4 mm wide and 3 cm long) form zones of multiple striping of this sequence and occasionally display bowed laminations that indicate the direction of intrusion. Thin-section examination shows structureless zones in overlying laminated quartzite leading into some of the dykes, which suggests a liquefaction origin as explained for similar features at locality 1. Several 5- to 10- cm long dykes, not quite in contact, are superimposed vertically above one another where the lower dykes possibly come from source beds now missing.

Locality 3 (Figure 2.8) is 900 m along the Serpentine Dam Road from its junction with the Gordon River Road. The outcrop is an alternation of layers of phyllite and quartzose phyllite containing approximately planar quartzite bodies similar to those described from locality 2. The average length is 5 cm (40609, 40610), though examples as long as 20 cm are known (40711); the average width is 3 to 4 mm. Their general morphology is similar to the 'molar tooth' structures described by Clarke (1970), except that here they are subparallel to a strong crenulation cleavage (52.185).

Besides the above localities which were sampled and studied in some detail, it has been found that virtually all zones of pelite outcrop contain clastic dykes. They are prominent in the pelites south of Mt. Sprent (7402014), from Greycap to The Citadel and around the Gordon Dam, particularly on the road to the Access Tunnel. Figure 2.17 shows specimens from the last mentioned locality where the dykes survive despite the presence of two well developed crenulation cleavages.

Many mudstone outcrops have extremely and irregularly lenticular sandstone bodies which do not possess the normally symmetry of boudins (Figure 2.12 b). Their isolated and irregular nature suggests they are infilled scours. Angular mudstone fragments occur from time to time in sandstone layers (Figure 2.18 a). The fragments are normally 1-4 cm in size but may reach 15 cm. Several explanations are possible for these features including an intra-formational breccia or a dessication origin. A very common feature of the quartzite layers, within the mudstone association, is the presence of variably elongate black blebs (Figure 2.18 b, 46272). Unfortunately when these specimens are sectioned, the contents of the blebs have invariably been weathered out. It is possible that these blebs were once carbonate concretions as described earlier, though it is possible they were mudstone fragments. This latter explanation is not favoured because of the close association of distinct mudstone flakes and the blebs in Figure 2.18 a. The thick (up to 2 m) sandstone layers of the mudstone association, occasionally show internal structures in the form of cross-bedding but on the whole are glassy in the field and totally recrystallised.

2.2: iii) <u>Meta-carbonate rocks.</u> The great bulk of the meta-sedimentary rocks of this region contain negligible amounts of calcium (see analyses Chapter 5). Some quartzite types do contain widely separated carbonate ellipsoids that may reach 6 mm in their longest dimension (46310). Many quartzite layers within the phyllite dominated horizons of the Gordon Dam area, contain small black blebs which are perhaps the deformed equivalents of carbonate ellipsoids. Surface outcrops of this rock type are now almost entirely leached of their carbonate component. Such occurrences, however, are infrequent in the region studied. Rarely, coarse siltstone layers, from clastic dyke locality 1, have iron-rich carbonate contents of up to 40% (40636, 40641, 40683, 40684). Although most of the amphibolite occurrences in the region can be demonstrated by their tectonics and field relationships to be igneous in origin, an amphibolite of unusual composition exposed in the Gordon Road some two kilometres east of locality 1 (Serpentine lookout), may be a metamorphosed clayey dolomite. This Gordon Road amphibolite is concordant with the general lithological layering and is extremely rich in MgO (Table 5.VI). The bulk composition appears to indicate an impure dolomite parent and a sedimentary origin is supported by a more complex textural history than the typical meta-igneous rock of the region. Available evidence places the igneous intrusions as post-D_1 but before the metamorphic peak which was pre- and syn-D_1. The Gordon Road concordant amphibolite has a nearly penetrative crenulation and a later weak crenulation thus demonstrating the presence of the regional S_1, S_2 and S_4 cleavages. Either the amphibolite is an early intrusive or a sedimentary layer. Analysis of major element compositions does, however, point in some respects to an igneous origin. Plots of Niggli *mg* versus *c* show a field where compositions between typical pelite and limestone or dolomite should fall (Leake, 1964). Specimen 46213 has the following Niggli values; al = 12.22, fm = 70, c = 15, alk = 3.4, mg = 0.73, and on an mg/c plot lies in the region of early differentiates or accumulates from mafic magmas. Considering the tie-lines of typical pelite-limestone and typical pelite-dolomite, 46213 lies on the opposite side of the dolomite line to the limestone field. The sedimentary parent was, therefore, not a pure dolomite/shale mixture. Major element compositions are of limited use and the values determined do not preclude a sedimentary origin. Of greater significance is any trend evident in plots such as mg versus c or al-alk versus c. Magmatic differentiation trends are at right angles to trends caused by mixing pelite and dolomite. Not enough enigmatic amphibolites are present in the Strathgordon region to allow such a trend to be studied. The limited occurrence on the Gordon Road may possibly be clearly distinguished as metasedimentary or meta-igneous by a more complete chemical analysis of the type carried out by Armbrustmacher and Simons (1977).

One unusual carbonate rock was found in the underground excavations for the Gordon Power station. It is apparently an offshoot of dyke rock (Glynn Roberts pers. comm.) and is approximately 40% calcite, 45% chlorite together with minor quartz and phengite. Dentric alteration and partial assimilation of pelitic country rock with subsequent metamorphism could account for the mineralogy and composition (table 5.VI, 39166, 46300).

2.3: Environmental Interpretation of the Meta-sediments

On a regional scale the metamorphosed Precambrian of Tasmania is associated with carbonate rocks (Hughes 1975). On the Lyell Highway, near Mt. Arrowsmith, dolomite, interfoliated with schist, contains oncolites and oolites, and shows a complex diagenetic history of dolomitisation and silicification. Though the sedimentary history of this section has not been studied in detail, a development similar to a modern intertidal to supratidal situation is suggested. The common association of quartz arenite and carbonate rocks with this type of history has been commented on many times (see Swett, Klein and Smit, 1971, p. 400) and may imply a tidal origin for the associated sandstone.

The textural properties of the sandstone are remarkably like sand from aeolian facies but the cross-bedding is of a sub-aqueous origin. Aeolian cross-stratification is typically large in size (1-12 m) though displaying a large range in size (Horne, 1971). The cross-bedding in the quartz arenite under consideration is always small in size and the rapid changes of current direction also suggest a non-aeolian origin. Besides the size of the sets of cross-bedding, the palaeocurrent patterns provide the best information on the environment. The most consistent feature of these patterns is a bimodal, bipolar element which is very commonly assumed to indicate tidal action (Banks, 1973). Other modes that appear may be related changeable wind-driven waves, storm surges, or semi-permanent currents resulting from prevailing winds or global differences in temperature and salinity of the oceans. The overall characters of the quartz arenite are quite similar to shallow marine shelf deposits (Pettijohn, Potter and Siever, 1973, Table 11-6). A major difficulty is that the environment postulated is the hardest to investigate in terms of recent analogues; access is restricted and present continental shelves are in disequilibrium. The nature of quartz arenite deposits such as those in southwest Tasmania, may, in large part, reflect conditions quite different from the present day.

Anderton (1976) has proposed a tidal shelf sedimentation model for the Jura Quartzite of the Late Precambrian, and in a detailed study was able to determine the relative importance of tidal and storm conditions. Many of the features described by Anderton are found in the quartz-arenite of the Franklands and comparisons can also be made with successions described by Banks (1973), Klein (1970), Swett et. al. (1971) and McCave (1973), all interpreted as being of a tidal origin. The influence of storms is difficult to ascertain; equivalents of Anderton's 'parallel laminated facies' are present and thus may reflect deposits from tidally - dispersed, storm generated suspension clouds. The definitive storm related type 3 erosion surfaces of Anderton have not been observed though this may be a function of lack of suitable outcrop. The environment envisaged is a shallow tidal sea bordered by exposed sandy beaches adjoining a hinterland where wind was a very important transporting agent.

Swett, Klein and Smit (1971) have noted the textural maturity of deposits with an inferred tidal history. They have estimated transport distances for quartz grains in such an environment and, though these seem to be overestimates for the majority of grains in the population, great distances are indicated. Kuenen (1960, p. 109), however, believed that only in the desert dune environment is sand effectively rounded and several other authors have suggested a marked aeolian phase in the history of quartz arenite. Pettijohn, Potter and Siever (1973) have noted that the bulk of quartz arenite was formed in the Late Precambrian and Early Palaeozoic. A major factor in this time concentration is likely to be the lack of vegetation such that the regolith was not bound in the same way as in the post-Devonian situations. Wind action would have been much more important with perhaps only algal mats in intertidal environments to act as cover to sedimentary particles. A similar conclusion has been reached by Chandhuri (1977) for a Middle Proterozoic sandstone. It seems reasonable, therefore, to suggest an aeolian source for the quartz arenite which was then extensively reworked in a tidal, shallow sea. Some objections to a tidal hypothesis may arise from the length of time required to produce each alternating set of cross-stratification. This is certainly much longer than each cycle of ebb and flood current. Caston and Stride (1970) have, however, measured a pattern of separation of the ebb and flood currents which if moved laterally with time could readily account for herring-bone, cross-lamination.

Tidal currents imply periodic motion caused primarily by the attraction of the Moon on the surface waters of the Earth. The history of the Earth/Moon system is a matter of some controversy and many authors believe that, because of their dissimilar bulk composition the two bodies have not always been closely associated. A hypothesis of particular note here is based upon prediction of behaviour during the slowing of the Earth's rotation. Direct extrapolations suggest that the Moon would have been about 20,000 km from the Earth's surface in the period 1,000 - 800 m.y. ago. Tidal energy would be increased by a factor of 8,000 creating huge tidal waves and partial or complete melting of the Earth (Tarling, 1975). Clearly the geological record does not substantiate such a prolonged catastrophic event though the concentration of quartz arenite has been related to greater tidal activity at that time. Klein (1970), however, has studied the frequency of fossil, clastic, tidally-dominated successions through time and concludes that they are concentrated in the Devonian and Carboniferous. Obviously more information is required to resolve the conflict. Considerations of planetary dynamics appear to be quite misleading and from geological evidence Truswell and Erikson (1975) and others have established tidal origins for Lower Proterozoic units, thus confirming the presence of the Moon more than 2,000 m.y. ago. Indeed catastrophic events such as those predicted by astronomical observations cannot have happened within the last 3,000 m.y. and given that differences such as the lack of vegetation have occurred, many Late Precambrian depositional environments can be explained in terms of models generally based on modern situations; drastic modifications to these models are not required.

The lithologies and assemblages of sedimentary structures of the mudstone association are strikingly similar to the *Diplocraterion yoyo* facies group of Goldring (1971) which is most likely to be delta-front platform (see also Coleman and Gagliano, 1965) or inner continental shelf deposit. Goldring (op. cit.) also considered a tidal-flat origin for the *D. yoyo* facies but this was mainly discounted because of the considered lack of graded beds in the modern equivalents. Anderton (1975) has proposed a tidal-flat and shallow marine origin for the Craignish Phyllites which have several features in common with the mudstone association. Grading in the Craignish Phyllites has been attributed to shallow marine, storm conditions (Reineck and Singh, 1972). In many situations trace fossils (Goldring, 1975) provide critical evidence as to the environment of formation of a particular deposit. Such features are lacking in the sections described here and interpretations rest heavily on structures and lithology.

The mudstone association appears to be dominated by the effects of pene-contemporaneous erosion and scours of similar size and cross-section occur as backwash channels on modern tidal flats. Channels of this nature with steep to overhanging banks have been observed on recent intertidal falts. Meandering channels are 'fair weather' tidal distributaries, straight channels are found to be parallel to wind directions during storms. Goldring (1971) considers that such features can occur from the intertidal zone down to depths of 3 m. The thicker quartzite layers within the mudstone association may be point bar deposits.

2.4: Environmental Conclusions

The mudstone association appears to have formed partly in an intertidal zone and partly in the subtidal zone immediately in front of a deltaic or tidal-flat environment. Precise analysis is restricted by the lack of information much of which may never be forthcoming because of the age of the deposits.

The arenite association material was derived from an environment lacking in vegetation which allowed wind erosion to predominate. This material was incorporated in the marine environment (by transgression?) and extensively reworked in a shallow-shelf sea under tidal influence. Ginsburg (1975, p. 92) has listed four categories of sedimentary structures that are indicative of a tidal-flat or tidalite origin. Evidence is required for (1) rapid current reversals, (2) alternating slack water, strong current, (3) intermittent subaerial exposure and (4) alternating erosion and deposition. The most important features of his categories 1, 2 and 4 are found in the Franklands. Features related to intermittent subaerial exposure (category 3) have been reported only a few miles northeast of the Gordon Dam in quartzite of the same age (McClenaghan, pers. comm.).

The extremely high percentage of non-undulatory, quartz-grains in the rare undeformed quartz-arenite layers indicates a source region of very high grade metamorphic rocks (granulite, migmatite) or felsic plutonic rocks (Basu et. al., 1975). Provenance studies based on heavy mineral assemblages can be quite misleading in rocks where there is evidence for high solution activity during diagenesis (Dana Rusell, 1937), and were therefore not attempted.

2.5: Meta-Igneous Rocks

Less than 1% of outcrop is of metamorphosed igneous rocks. In natural exposures discordant relationships are difficult to prove and contact relations are usually obscured. Many bodies transgressive to the sedimentary layering have been mapped in underground working of the Hydro-Electric Commission (Andric, Roberts and Tarvydas, 1976; Roberts and Andric, 1975). Several of these dykes display good preservation of igneous textures and in rare cases igneous mineralogy is seen in various stages of transition to metamorphic assemblages. Only one specimen of a dyke rock was chemically analysed and this showed a tholeiitic basalt composition (Table 5.VI). In road cuts around the Gordon Dam several transecting bodies are exposed at the surface. On the bench to the west of the Busbar Shaft, an amphibolite dyke is slightly oblique to the general dip and strike of the compositional layering and the fold hinge lines. Several amphibolites clearly cross cut bedding in massive quartzite along the access road to the base of the Gordon Dam (now under water).

CHAPTER THREE

MINOR STRUCTURES, MICROFABRIC AND STRUCTURAL SEQUENCE

3.1: Establishment of Structural Sequence

Overprinting relationships have provided a firm basis for the determination of the sequence of deformational events in the Franklands/Wilmots area. In key outcrops structures of four phases are recognised and within some small areas (10's of square metres) five distinct cleavage producing events may be proved. Minor structures of all events are markedly heterogeneous in their development, not just from one lithology to another but within the same rock type. As a result, individual outcrops commonly contain evidence for two deformational events and some for only one or they may be undeformed. If an outcrop displays two interacting sets of structures, they can be placed in the regional sequence in a variety of ways which all depend upon extension of certain attributes from key localities. Many authors have considered the difficulties of structural analysis in regions of discontinuous outcrop (reviewed by Park, 1969 and Williams, P.F., 1970). Problems are alleviated by the good exposure in the mountainous areas and the excellent continuous exposures in the Gordon and Serpentine Dam areas. Information from the latter region has proved critical in that overprinting is common and a wide range of lithologies are present. Another helpful factor, on a regional scale, has been the constant attitude of structural units (enveloping surfaces to 2nd order folds) which can be followed along strike from areas of dominant D_2 effect to D_3, etc. Transition have not always been sampled but the general uniform attitude of the major units allows two sections to be superimposed and the minor structures to be placed in the overall sequence. An example of this procedure is found in the comparison of the eastern slopes of the Central Wilmots and the Southern Wilmots (Figure 3.1). In the former area bedding dips regularly between 50 and 70° to the east and a penetrative cleavage (S_2, transecting D_1 isoclines and boudins) dips moderately east. Along strike where bedding maintains a constant attitude, a steep, west-dipping crenulation (S_3) predominates and in both areas a near vertical, slightly easterly inclined crenulation (S_4) overprints all earlier surfaces. The post-D_1, pre-D_4 surfaces rarely interact and the S_3 is best placed in the sequence by its similarity in orientation to the proved S_3 in similarly dipping quartzites on The Starfish. Time relations are confirmed in the section between Mt. Sprent and the Serpentine Dam (Figure 3.2).

Correlation problems in this region are complicated by the very similar strike bearings in each sub-area, of all generated structural surfaces. Attitude alone is a poor criterion for correlation as early surfaces are disoriented and later ones are deflected at structural heterogeneities and lithological contacts. Overlap of attitudes is extensive and the general, near coaxial nature of the folding further increases the possibility of placing structures in the wrong deformation event when using orientation alone. In the majority of areas studied, S_1 is often found parallel to bedding and the pre-D_2 geometry must have involved regular piles of isoclines. D_2 and D_3 folds are rarely major folds and hence a regular bedding attitude often defines the enveloping surface of the pre-D_4 situation. The post-D_1 rotation of large thicknesses of material appears to be mainly D_4 in age and S_2 and S_3 were involved in these rotations.

A particular problem in field analysis is the similar outcrop expression of S_1, S_2, S_3 and S_4 in quartzite. Thin section analysis may resolve the difficulty but is not always successful particularly in pure quartzite. A related feature is that D_1 to D_4 folds may all appear to have bedding as the form surface. The D_1 fabric may be so weak that it is unrecognisable or later events are sufficiently strong to overprint S_1 and the potential for error is, therefore, introduced. A further complication arises from the simulation of tectonic folds by primary deformation of cross-bedding which may lead to a greater degree of complexity of tectonic sequence than actually exists.

Phyllite exposures are often extensively crenulated and where overprinting occurs, the crenulation cleavages produced in D_2 to D_5 can have identical morphologies, though S_2 is usually the most differentiated surface. This latter feature may be used carefully as an aid in correlation, but it appears that all S_2 crenulated phyllite is not well differentiated. Despite the similarity of crenulations from several events, phyllite exposures are very useful in determining the sequence of events. Many outcrops in this rock type have evidence for three phases of deformation and within small areas further distinct phases may be defined. A common occurrence involves folded quartzite layers with $S_0//S_1$ and a generated S_2 cleavage with both the F_2 axial surface and S_2 being folded by F_4 and transected by S_4 (Figure 3.8 c). The regionally developed S_3 is very patchily developed in the phyllite units. The advantage of good control in phyllite is negated, to some extent, when correlations are made across lithological contacts. Well developed crenulations in phyllite may disappear in micaceous quartzite or quartzite. Refraction and changes in morphology are major problems with lithological variation.

In areas of more discontinuous outcrop, folds, cleavages, lineations, etc., may have to be assigned to a particular structural event with some uncertainty. Several examples have not been placed in the sequence when extrapolation appeared unjustified. The degree of exposure is usually sufficient to allow the monitoring of variations related to lithology and metamorphic grade and, combined with frequent examples of overprinting and extensive areas of regular interaction of phases, the overall structural sequence can be established with confidence.

3.2: Minor Folds and Boudinage of the First Deformation Event

i) Quartzite structures. Minor folds produced in D_1 with bedding as the form surface are well represented throughout the quartzite outcrops. Away from the eastern Franklands, the majority of these folds are very tight to isoclinal with long limbs with no indication of vergence. F_1 of this nature are very strongly flattened (Figures 3.3a and 3.4b) with profiles approaching class II (Ramsay, 1967). A superimposed flattening of between 1 and 20% is described as low; 21-50% moderate; 51-70% high; 71-99% strong. Asymmetric coupled isoclinal folds are sufficiently common to allow the definition of major D_1 folds. Individual folds within the couples vary from open with sub-angular closures (Figure 3.5 a) to isoclinal with rounded closures (Figure 3.5 b). In all cases the common limb is thickened with respect to the external limbs and the thickening presumably occurs by flattening whilst the short limb is at a high angle to the overall layering. The variation in final attitude of the common limb must depend on how far the deformation continues to rotate the limb. Minor D_1 folds occur with the greatest frequency in a lithology that involves quartzite laminated on a scale of less than one centimetre in thickness. The finely laminated material often occurs in groupings up to 10 cm thick and may involve thin micaceous partings. Within such alternations the prominent, thick quartz arenite layers are allowed a degree of independent behaviour and the small folds are disharmonic. Pure quartz arenite layers are normally within the 10 - 30 cm thickness range but extend to a maximum of 2 metres. This lithological association predominates in the Eastern Franklands where F_1 structures are consequently abundant. A general correlation can be made between the type of multilayer affected and the overall nature and geometry of D_1 folds (cf. upper and lower parts of Figure 3.3 b). Thickly banded units have rounded hinge zones and profile geometries approximating those of parallel folds (class IB) flattened less than 30%.

A further multilayer type is well represented on the lower slopes of Murphys Bluff and Cleft Peak where regular bedding is within the 10 - 20 cm thickness range but no thinly laminated portions are present. In this zone, piles of slightly vergent, near chevron, tight folds (Figure 3.6) are well exposed on joint faces up to 150 m in vertical extent. Flattening values in the folds are low to moderate but the nature of the outcrop precludes a detailed examination. Despite their extensive development, the folds do not carry an axial plane foliation (S_1) except in very rare situations.

The orientation of D_1 folds varies considerably due to rotation in later deformations. Axial surfaces vary from vertical to flat lying and hinges vary in bearing from a little east of north anticlockwise to a little north of east. Plunges are generally gentle to moderate to the north or west but there is some evidence for a spread of hinges within the layering in small areas. The latter feature may be a result of variable amounts of stretching in the D_1

strain ellipsoid (Sanderson, 1973).

Boudinage is very commonly associated with minor D_1 folds in quartzite and appears to depend on similar lithologies for formation. Boudins are normally best displayed in sequences with layers of marked lithological and competence contrasts but this does not appear to be the case in the majority of examples seen in quartzite multilayers of the Franklands. Figures 3.6, 3.5a, 3.5b show boudinage which in the first illustration is in pure quartzite and in the latter outcrops involves alternations of finely laminated quartzite with thicker beds. Boudin profiles are usually rounded and regular pinch and swell structures (Ramsay 1967) are developed in the evenly laminated multilayers (Figure 4.41a). Such features suggest a low ductility contrast between layers in the sequences and the presence of boudins indicates that the strain rate under the prevailing P/T conditions did not allow the quartzite to behave in a completely ductile manner. Boudins do, however, occur in quartzite with intense microfabrics which appear to have undergone extreme strain. Boudinage would most likely have occurred late in D_1 in this material.

3.2: ii) Interlayered Quartzite/Phyllite structures. D_1 folds are exceptionally rare in units where phyllite is an important component. When recognised they have extreme flattening values and hinge bed thickness to limb ratios exceed 15 to 1 (Figure 3.7 a). Some minor D_1 folds approach a 'forked layering' appearance and many examples of this feature cannot be resolved into a tectonic or sedimentary category. The presence of small blebs of material within quartzite layers gives a qualitative indication of strain in some of the minor D_1 folds (Figure 3.7 b). The pattern of the markers is very similar to that shown by Ramsay (1976, Figure 16c) where high ratios occur on the inner arc and the ratios lessen towards the outer arc and a slight fanning is present. Ramsay (op. cit.) considers this pattern to be indicative of a low competence contrast during deformation with high initial values of layer parallel shortening. All the minor structures of the quartzite/phyllite assemblage indicate very high strain. The quartzite layers possess a penetrative fabric and are totally recrystallised. Minor folds and bleb markers are intensely flattened and clastic dykes where $S_1//S_0$ are shortened considerably.

Boudinage is very common in quartzite layers surrounded by phyllite but its recognition is clouded by the tendency for later folds to be localised at zones thinned in D_1 (Figure 3.8 e). Also sedimentary scour and fill structures may simulate boudins (Figure 2.12 b and 3.7 a) especially if internal lamination is obscured by recrystallisation. The tectonic structures occasionally contain the qualitative markers seen in some early folds. Where D_1 effects can be separated from those of later deformations and studied along, a rapid increase in strain is observed towards the neck of the boudin. The nature of the failure is clearly ductile.

Refolding of minor D_1 fold axial surfaces by F_2 or F_3 is not seen in the interlayered quartzite/phyllite. Occasionally F_1 and F_2 folds are found centimetres apart with opposite senses of asymmetry affecting one layer (Figure 3.9).

3.3: D_1 Microfabric

An attempt has been made at isolating D_1 events by searching for specimens lacking obvious polyphase effects on the microscopic scale. The study is complicated by the post-D_1 metamorphic climax which may have induced recrystallisation by static annealing. To minimise the effect of the later thermal event, a region of low metamorphic grade was chosen where pelite intercalations tend to be slaty rather than phyllitic or schistose. A sequence from very low to moderate strain will be examined where a modification of the method used by Elliott (1970) is employed to quantify tectonic strain indicated by quartz grain 'shape factors' (Chapter 7 reports the methodology in detail). This sequence of increasing strain will be used to relate intensity of deformation to the degree of development of optical strain effects and percentage recrystallisation, thus suggesting the dominance of dynamic recovery over static (White, 1973).

Some material has survived to the present day in what must be nearly its pre-tectonic condition (e.g. 46295). Detrital quartz grains show only very slight undulose extinction and are set in a binding agent of virtually pure quartz overgrowths. No recrystallisation is visible in this very lowly strained material. The first sign of the operation of dislocation deformation mechanisms is the production of sub-grains entirely within pure quartz overgrowths (44376, Figure 3.10). Though this specimen was measured as being unstrained, the accuracy of the method would allow a few percent shortening in the Z direction. A very low strain is indicated. At 15% shortening in Z, the detrital grains are somewhat more undulose and small, distinct new-grains are seen in the quartz overgrowths (44373). Recrystallisation is usually concentrated on the boundary between the detrital grain and the overgrowth; the preservation of dust coatings on the grains demonstrates that the recrystallisation occurs only in the overgrowth which is approximately 25% new grains. The next stage in the sequence is represented by a specimen (44375, 19% shortening in Z) with a slight increase in mica content. Very little overgrowth (10% approximately) is preserved in the original form and is now represented by small new quartz grains and perfectly aligned platelets of mica. Even in mica free zones, the new quartz grains show a good dimensional preferred orientation which is strongest in the XY plane. Very pure specimens with 28% shortening in Z also show a nearly total recrystallisation of the overgrowths and presumably this state is achieved in the purest arenite at about 20 to 25% strain.

Very similar localisation of recrystallisation in quartz overgrowths has been figured by Majoribanks (1976, Figure 6) but he is unclear in the text as to origin of the "fine matrix grains". The implication is that these grains are detrital but have suffered more deformation than the larger grains and hence are more recrystallised. Apparently in the case of Marjoribanks' study (op. cit.) both fine detailed grains and overgrowths may be present leading to uncertainty. Similar mixtures have not been observed in the arenites now under discussion. It is believed that the differential degree of recrystallisation of overgrowth before detrital grains is evidence for hydrolytic weakening (Griggs, 1967). Jones (1975) has recently demonstrated plastic deformation of quartz in fibrous veins at temperatures around 250°C. The quartz in these tectonic veins was deposited from solution and has very high hydrogen contents; the overgrowths have been formed in a similar way and, therefore, would be expected to have similar strengths. Detrital grains of the Franklands' arenite are probably from a dry, high-grade metamorphic terrain. Though the measurement of hydrogen content by Jones (op. cit.) may not be accurate (Tsong, McLaren and Hobbs, 1976), all studies to date have used the infrared absorption spectroscopy method and thus relative orders of hydrogen contents are maintained. The mechanism of hydrolytic weakening is still a matter of debate but several studies have demonstrated the reduction of strength in quartz brought about by hydrogen (Bell and Etheridge, 1976; Morrison-Smith, Paterson and

Hobbs, 1976). Conditions during D_1 are uncertain but from pelite assemblages in the eastern Franklands (quartz-phengite-chlorite without garnet or albite) temperatures are not expected to have exceeded 400°C. At the initiation of D_1, temperatures of the order of 200°C are likely to have operated and load pressure would have increased in the first event by tectonic thickening during nappe formation as proposed by Dewey and Pankhurst (1970) for the Caledonides. The temperature may, in fact, be indicated by the behaviour of the overgrowths, as Jones (op. cit.) has shown experimentally that the concentration of hydroxyl ions in quartz falls off rapidly above 300°C. Above this temperature, dehydration may be expected to level the hydrogen concentration in the overgrowths and detrital grains and not allow such disparities in mechanical behaviour. The degree of strain heterogeneity between overgrowth and detrital grains is considered to be small because of the grain supported nature of quartz arenite. Stress can, therefore, be transmitted through the grains and the very marked differences in recrystallisation must have occurred when the grains and overgrowth had undergone similar amounts of change and shape.

Specimen 44375 at moderate strain (Z = 0.81 r) showed near total recrystallisation of the overgrowth but contained slightly higher traces of mica than usual. The mica is presumed to have been a minor argillaceous component of the original overgrowth (Majoribanks, op. cit.) which on crystallisation released water to further weaken the quartz overgrowth and enhance recrystallisation. The mica platelets appear to have exercised a control on the growth of the new quartz grains but in mica free areas these grains are still aligned. Even in a specimen with less than 20% shortening in Z (44373, Z = 0.85 r) the mica platelets are perfectly aligned, arguing against mechanical rotation of pre-existing grains and favouring recrystallisation, in a direction controlled by stress difference. Well before a dimensional preferred orientation of detrital quartz grains is a marked feature of an arenite, the recrystallised overgrowth (including minor mica) gives a tectonic grain to the rock (Figure 3.11). Such a grain is detectable in the field and is the first expression of deformation.

At 20% shortening (44375) sub-grain development is noted in approximately 25% of detrital grains. Deformation lamellae and broad hazy deformation bands are not common and the latter may be extreme examples of undulosity. No distinct new grain growth in the detrital grains can be detected. When the 28% shortening stage is reached (44372), the grains are more strongly undulose (c-axis mismatch within individual grains increases), deformation lamellae are fairly common and extremely limited new grain growth has occurred in the detrital grains. Thin deformation bands (0.05 mm) with fairly sharp boundaries (Figure 3.12 b) are also often noted with a thickness 5 to 10 times that of deformation lamellae. Within grains the trace of deformation lamellae is often curved or kinked with waves of extinction running approximately parallel to the axial planes of these features. In the low strain states undulose extinction is often concentrated at the contacts between detrital grains and may radiate away from such points. The next sample in the strain sequence, 44374, gives a shortening of 36% and the increment from 28% to 36% involves quite a marked change in the micro-fabric. Approximately 30% of the detrital grains are recrystallised and there is a wide range from very slightly undulose grains to those almost entirely consisting of fine new grains. Some of the partly recrystallised grains are a chequerboard of two, high-angle, sets of fine (0.05 mm) deformation bands which appear to grade into zones of new grain growth (Figure 3.13 a). Away from the grain growth, the fine deformation bands are rectangular areas with different lattice orientations, very similar in form to the single sets seen in 44372. One direction is parallel to faintly preserved deformation lamellae perhaps supporting the contention of Gay (1974) that a high degree of lattice rotation along lamellae, together with boundary migration, creates deformation bands. The recrystallisation of the band boundaries accords with this process being favoured in areas of highest misorientation in old grains (White, 1976). In other grains (averaging 1 mm in size) large subgrains have developed into 4 to 6 new grains and smaller new grains line their boundaries. Typically subgrain within subgrain structure is seen and if the small subgrains are sufficiently rotated to form new grains then recrystallisation will also be localised on their boundaries to give a nearly complete fine recrystallisation of large original grains (Figure 3.14). This sequence of polygonisation (Figure 3.13b) and then new grain growth along boundaries of misorientation is the most common path of recrystallisation. Several examples are, however, noted of normal sized (0.1 to 0.2 mm) deformation bands in 44374. Along the deformation band boundaries all stages from faint subgrains to distinct new grains occur and such zones are sites of recrystallisation (Figure 3.15; see Bell and Etheridge, 1976). Subgrains also occur in the deformation bands. Uncertainty exists as to whether the wide deformation bands form by boundary migration of the fine types or whether the large type initiates with a greater boundary spacing. The latter is favoured because of the bimodal distribution of band widths, with no intermediates being recognised. A small proportion of highly recrystallised grains, preserve elongate and aligned portions of the original grains, illustrating an early deformation band development. Though optical features indicate a heterogeneous strain on the scale of individual grains, the mantle/core structure emphasised by several authors (e.g. White, 1976) is quite rare. A variation of this type is seen between diagenetic overgrowth and detrital grain but not within initially homogeneous grains.

In D_1, the evidence for diffusion or solution transport (Durney, 1972) is limited. Syn-tectonic, fibrous, extensional veins are found (44372) but not frequently. Quartz and mica beards on detrital grains can often be demonstrated to be in situ recrystallisation of overgrowths. In examples of matrix rather than cement recrystallisation, Williams (1972) has suggested that quartz and mica beards do not depend upon growth in 'pressure shadows' though some small component of this nature may be involved. As detrital grain boundaries are usually preserved in the quartz arenite, the possible action of diffusion flow is restricted to the binding agent except for the high strain zones of 44366. Sutured, stylolitic boundaries between detrital grains are found here and they appear to be syn D_1 (see strain analysis chapter for photographs). 44366 is the only specimen to show this behaviour, which may be related to work hardening necessitating a change of deformation mechanism. This suggestion is supported by the observation that dynamic recovery processes appear to be less effective than in other studies (e.g. Bell and Etheridge, 1976) as deformation lamellae are often preserved.

Complete gradations are known from quartzite with about 50% new grains to 95% new grains and very few remnant old grains (46264 to 46275 to 46263 to 46279). Unfortunately in each example it is impossible to be certain of the initial nature of the rock, for, despite their quartz rich nature, they may have been in part finer grained. Later deformations may also have been responsible for some new grain production. Therefore it is uncertain whether the fine grainsize can be ascribed solely to tectonic processes. Within the progression towards the thrust of the Franklands (the strain sequence outlined), no more than 50% recrystallisation is found. Some pure quartzite layers are completely recrystallised and have very strong D_1 fabrics (46210, 46215). New quartz grains in these specimens are about 0.03 mm in size; they are inequant and define S_1 in association with tiny mica platelets. The purity of these rocks may indicate derivation from the typical arenite of 1 mm to 2 mm grainsize, which suggests that the end product may be interpreted as resembling many mylonites in texture, but not in the overall structural relationships (Bell and Etheridge, 1973).

In quartzose phyllite and phyllite, S_1 is a pervasive fabric dominated by a perfect alignment of mica flakes. The quartz grains between the micas are small, usually less than 0.05 mm, and have varying degrees of dimensional preferred orientation. Specimens little affected by post-D_1 effects or in microlithons of S_2 etc., tend to have near equidimensional polygonal fine grains. Such materials do not show any remnant sedimentary textures.

3.4: Minor Structures of the Second Deformation Event

i) Quartzite structures. Within quartz arenite assemblages the D_2 folds are somewhat rare but are easily recognised because bedding or $S_1//S_0$ has undergone rotation in discrete zones thus showing similarities with kink bands (Figure 3.16). The rotated limbs show layer thickness variations which show a direct correspondence with the angle of external rotation and the angle between the kink plane and the external foliation (Anderson, 1964). The illustrations depicting these folds show rotation arrested at various stages with resulting style differences (Figure 3.16). On the thinned limbs, sutured layer interfaces (Figure 3.16 a) suggest some dissolution took place to accommodate the rotation though crystal plastic processes were probably more important. All examples in the kink category are negative according to the classification of Dewey (1969). Poorly defined D_2 folds occur in strongly boudinaged, arenite multi-layers where thinned portions show a consistent sense of rotation. Best development occurs where necks were stacked one on top of another.

High values of common limb attenuation and rotation may produce folds that have morphologies similar to D_1 folds. Detailed analysis of the associated foliation and form surface is necessary to place a particular fold into the structural sequence, particularly in isolated outcrops. This problem is emphasised by the presence of kink style D_1 folds in the massive quartzite of the Gordon Dam site. The D_2 kinks have a remarkably consistent sense of asymmetry in a region from Frankland Saddle to the lower slopes of the Southern Wilmots. One outcrop shows a possible opposite sense of vergence (Figure 3.16 g) but this example is not unequivocal as it may be the interference of an F_2 with an F_1.

The characteristics of D_2 folds are remarkably sensitive to the nature of the affected multilayer. Along the backbone of the Wilmots, F_2 structures are common in quartz arenite that contains thinly laminated intercalations of quartzite. Here D_2 folds vary from open to tight and have rounded closures with low to moderate degrees of flattening. Generated axial plane structures are well developed but were lacking in the kink structures.

Orientations of D_2 folds are again very variable due to rotation in later events and show the same variation noted for D_1 folds.

3.4: ii) <u>Interlayered Quartzite/Phyllite Structures</u>. Where phyllite exceeds 20% of a unit, D_2 folds, defined by pure quartzite layers, are quite common. The majority of the minor folds are very tight or isoclinal and are well flattened (Figure 3.8 b, c, d). Rigorous geometric analysis is limited because the layers folded during D_2 already had considerable thickness variation due to D_1. Such variations controlled the development of minor D_2 folds as the thinner parts of a layer were folded in preference to the thicker zones. Powell (1969a) included the presently grouped D_2 folds in his first deformation and concluded that his D_1 could have been initiated whilst the sediments were not completely lithified because of the irregular layer thickness changes. The correct identification of D_2 folds and the early history of the layers obviates the necessity for soft sediment deformation.

The amount of strain in quartzite layers associated with D_2 is difficult to estimate. D_1 necks have been folded and displaced some distance from their original position (Figure 3.8 e). The accentuated fabric towards this type of F_2 closure may be partly a result of D_2 strain (Figure 3.17 a) but similar features appear in little modified D_1 boudin necks. Well developed fissility caused by $S_1//S_0$ is locally more apparent in D_2 fold closures and under certain conditions S_1 may be enhanced by D_2. An alternative suggestion is that D_2 structures are localised in zones of high D_1 strain.

3.5: D_2 Microfabric

In D_2, D_3 and D_4, the microscopic products of deformation show evidence for the operation of diffusive mass transfer, intracrystalline dislocation and microfolding processes to varying degrees. Some materials show evidence for all three modes of deformation, whilst in others, one may have operated to the exclusion of the remaining two mechanisms. D_2 displays this range particularly well and is noteable for the extreme examples of metamorphic differentiation produced in slightly micaceous quartzite and in phyllite. A prominent feature of many phyllite outcrops, is a well differentiated crenulation cleavage (Rickard, 1961), producing a tectonic striping with mica rich layers up to 4 mm thick. This fabric is best developed in laminated quartzite and phyllite (1-2 mm thick) and depends upon local heterogeneity for its formation. S_2 in phyllite thin sections is often quite variably developed spatially (41705, 40219) and in terms of degree of differentiation (Figures 3.18 and 3.19). The crumpled S_1 fabric within the microlithons is often strongly truncated against the differentiated zones and if a deflection into these areas is present, it is usually over a small distance relative to the lithon thickness. The differentiated layers are nearly pure mica in the majority of specimens even in the thinnest examples where this type of feature is first recognised. Remnant quartz grains tend to be much smaller within the zones than in the microlithons (46216) and examination of polished thin sections, reveals that haematite enriches these zones (Figure 3.20 a). Tourmaline and zircon follows this enrichment pattern but in an irregular fashion because of widely differing initial concentrations from specimen to specimen. Detailed microprobe work carried out by Boulter and Raheim (1974) showed that S_1 phengitic micas had a uniform composition. Within the S_2 differentiated zone, cores of phengites sometimes preserved this original composition but generally speaking the mica along S_2 was completely readjusted to a new composition. The chemical differences though slight, are significant. From S_1 micas to S_2, weight percent SiO_2 and K_2O are consistently decreased whilst TiO_2 increases.

Several different views have been expressed on the mechanism of formation of crenulation cleavages and the debate has not been simplified by the confusing terminology that developed around the subject. Cosgrove's (1976) recent contribution, in association with studies by Williams (1972b) and Means and Williams (1972), has led to considerable advances and clarification in the analysis of crenulation cleavage. The large variety of crenulation cleavage types are formed by the same process, involving microfolding which initiates stress gradients which induce the redistribution of minerals (Cosgrove, op. cit). Metamorphic differentiation is proposed as being an essential step in the production of crenulation cleavage and many authors have supported passive enrichment of mica as the dominant change (see Means and Williams, op. cit.). Others have proposed that the components of micas migrate towards the differentiated layers (Rast, 1965). Passive enrichment of mica is thought to be brought about by pressure dissolution, migration and redeposition of quartz. There is now a considerable body of textural evidence to support the operation of this mechanism and recent theoretical papers have given the foundation for a better understanding of the process (Rutter, 1976 and Durney, 1976). Petrographic evidence for S_2 shows corroded quartz and a concentration of opaques demonstrating dissolution but the role of the mica component in migration is difficult to assess. Partial preservation of D_1 chemistry demonstrates the rotation of the S_1 fabric and the small chemical changes generally registered indicate mobility of some elements. Without a known pre-crenulation or un-crenulated reference the sense of movement of components is uncertain. Unfortunately, a progression along a uniform layer from crenulated to uncrenulated material cannot be followed. A further unknown is the site of quartz redeposition: is it a cleavage/microlithon adjustment or does the quartz travel to a fracture and form a quartz vein? Many quartz veins are clearly pre-D_4 and post-D_1 but the variable development of S_2 precludes a time relationship to D_2 being established. In relation to mica chemistry (Boulter and Raheim, 1974, Table 1), the analyses show that the S_1 composition is only modified in structural zones that involve microfolding. The chemistry within these zones may reflect compositions required for equilibrium under the D_2 metamorphic conditions and that the structural processes merely provided sufficient activation energy to catalyse the readjustment locally. Etheridge and Hobbs (1974) have shown the need for caution in interpretation when analysing structural/metamorphic interactions. Cosgrove (op. cit.) considers the action of stress gradients in microfolding led to the migration of mica components both from the hinge to the limbs and vice versa in examples from Anglesey. Crenulations are invariably superimposed on an earlier fabric and metamorphic conditions can often be demonstrated to be different during the two events and the influence of both factors has to be taken into account.

The early formation of thin pure mica bands parallel to S_2 may suggest that the widening of the differentiated layers may occur by dissolution of quartz and incorporation of mica whilst S_1 maintains a high angle to the layer. Local anistropies may buckle S_1 close to the layers to give rise to a microfolding effect. This method, if important, must involve solution and recrystallisation of mica, similar to the mechanism of Holeywell and Tullis (1975), to explain mica preferred orientation in slates.

S_2 is associated with the production of a metamorphic layering in slightly micaceous quartzite, though microfolding features are not readily apparent in this rock type. Though S_2 of this nature is often at a low angle to S_1, a wide variety of angular relations are known. Two different morphologies which occur in this category are believed to be the result of basically the same deformation mechanism but certain characteristics complicate comparisons. The most commonly met expression of this microfabric (46246, 46248, 46249) is a series of discontinuous dark seams usually two or three grain diameters in length and spaced according to grain width (Figure 3.21 a). The seams are extremely thin (0.005 mm) and tend to anastomose a little around detrital grains, though this effect may, in fact, be due to two intersecting sets of seams or later folding. Iron oxides and traces of mica are the dominant components and the seams are slightly pleochroic indicating the presence of aligned, phengitic-mica. In the finer grained rocks the seams appear continuous as a result of coalesence of the more discontinuous types seen in coarser grained equivalents. Detailed examination shows the seams to be sutured and to have local large changes in orientation. It is this type of fabric, which despite an often inocuous microscopic appearance, produces a very pronounced 'grain' in weathered outcrop. Very thin films of interconnecting mica and opaques give rise to planes of weaker cohesion than well developed quartz alignement fabrics, presumably because weathering processes etch out the micas. Some correlation is noted between intensity of D_1 fabrics and that of D_2, but S_2 seam cleavages are developed in some specimens with no recognisable S_1 (44376).

In specimens where detrital grain boundaries are visible, the S_2 fine seam cleavages truncate the expected ellipsoidal outlines to varying degrees (Figure 3.21 a). This feature, taken with the composition, is best accounted for by a dissolution type of mechanism (Durney, 1972 and Elliott, 1973). Structures of this nature are considered to be sub-perpendicular to σ_1 (the maximum within the principal stress) rock, but are markedly influenced by local heterogeneities. The general theory proposes that a thin film of fluid occurs along grain boundaries. Paterson (1973) has demonstrated that the chemical potential of a mineral in a surrounding solution is directly related to the normal stress acting on its surface. The normal stress varies over the surface of a grain (Durney, 1976) and thus instabilities are formed which can be reduced by dissolution at the point of high normal stress. In the Franklands material under discussion it appears that the most common fabric prior to D_2 was composed of oriented ellipsoidal, quartz grains, partly recrystallised and set in a 'matrix' of totally recrystallised original cement. Because the S_2 films and the S_1 fabric are usually at a moderate to low angle, the highest normal stress would be expected to act along the flat face of the ellipsoid somewhat displaced towards one end of the grain. Dissolution would be initiated here and propagate laterally along the grain boundary. An interesting problem is the extension of the dissolution seam into the recrystallised overgrowth. Grain size is an important parameter in diffusive mass transfer flow processes. This was indicated by White (1976) in the rapid expansion of the Coble creep field in his deformation mechanism maps of quartz with decreasing grain size. Though pressure dissolution involves a pore fluid, similar controls are thought to apply and it may be expected that the process should be more important in the finely recrystallised overgrowth. The relative intensities of dissolution are, however, difficult to assess as the overgrowth invariably contained some mica and, in equal volume loss, the matrix would produce more insoluble residue. Stress trajectories about an ellipsoid may best explain the propagation of the dissolution planes (Cosgrove, 1976).

The second type of metamorphic layering in this group might be an accentuation of the first but certain features reduce the certainty of the correlation. In this case very pure mica bands up to 0.2 mm in width (Figure 3.21 b) are found at irregular intervals through several specimens (44376, 46262, 44368, 46265, 46254). Very similar features have recently been described by Harris et. al. (1976) but in more micaceous rocks. Groups of two or three bands with 0.5 cm spacing may be found with several centimeters between clusters. Specimen 46254 is unusual for the regular, closely-spaced, mica bands which occur every two or three grain diameters. Lengths are variable from 0.5 cm to 5 cm, though some may be traced discontinuously over tens of centimetres. In outcrop their irregular distribution, often at low angles to layering (44376, 46265), prevents ready recognition, though on broken surfaces the mica films are distinct. Saw cuts are often required to reveal the presence of the mica bands and only limited sampling has been carried out. A tectonic origin is suggested by their common inclination to bedding and markedly different composition from any sedimentary layer in the specimens studied. Dissolution of quartz and its migration away from the zones is considered the best mechanism to explain their features. Considerable grain shape modification is seen adjacent to the zones and is particularly demonstrable in fabrics with low D_1 strain. Here grains are of low axial ratios and grain shapes next to the mica bands are truncated half-moon forms and often scalloped (Figure 3.21 b). Within the zones, rare, very thin, quartz grains are the remnants of dissolution. Cataclastic granulation may give rise to grains of this type but the remnant grains are too small in abundance to be broken fragments of all the modified quartz grains (Figure 3.21 b). Granulation would also be a poor mechanism to account for the marked change in composition of the layers which may be pure mica in virtually pure quartz arenite (44376). Towards the thinner portions of the mica seams they take on a stylolitic, indented form and at their terminations bifurcate into several thin seams of the first group. Between the latter seams, are found very elongate grains that have an area much smaller than the average detrital grain away from the zones. This relationship suggests that the dissolved quartz was not redeposited on the low normal stress portion of the same grain but moved a greater distance. Away from the mica bands several specimens show a weak development of the first type of fine dissolution seams, sub-parallel to the bands.

Some of the bands appear to pass laterally into tabular zones where crystal plastic deformation processes were dominant (Figure 3.22 a). The transition appears to involve a small change in orientation of the two types of tectonic zones in some examples, though later deformations obscure this relationship. The plastically deformed zones have a quartz grain alignment oblique to their margins and are normally well recrystallised though hazy remnants of much elongated grains are preserved. Grains on the margin of the zone often appear to tail off into the recrystallised area (Figure 3.22 a). In specimen 44376, a finely recrystallised zone is bounded on either side by seam type cleavages. Though gradations across these features are rapid, they have the appearance of small mylonite belts and like their larger counterparts, some uncertainty exists as to the mode of formation. They may be zones of high strain caused by strong axial shortening at right angles to the belts with simple shear of the bounding blocks late in the deformation. Alternatively simple shear of rigid blocks may be the dominant external control throughout the history of formation. The relationship between the thin mylonite bands and those of mica appears not to be a direct one and the latter may have formed first and initiated simple shear movements because they acted as planes of weakness.

In the Mt. Sprent region and south along the Wilmots, S_2 reaches its peak development as a recognisable fabric in the field. The second cleavage is quite apparent in slightly - micaceous quartzite where the mica component is a little less or approximately equal to that of specimens showing the fine seam type of S_2. On the Wilmot Range this rock type predominates and shows evidence of derivation from quite coarse-grained quartz arenite in part. Specimen 46292 illustrates the S_2 microfabric particularly well as recrystallisation related to D_2 is well advanced. In this specimen the S_1 fabric is defined by hazy elongate grains usually between 0.3 to 0.5 mm long and 0.1 to 0.2 mm wide. These grains have developed subgrains and often new grains are seen within elongate remnants of grains deformed in D_1. The subgrains and new grains have a very strong dimensional preferred orientation at high angles to S_1 and have axial ratios of about 2 to 1. Within the S_1 grains, new grain boundaries are highly indented to define the S_2 direction. Very thin (0.01 mm) discontinuous lines of mica up to 1.5 mm long occur sporadically through the specimen sub-parallel to S_2 and very fine flakes parallel S_1. The dominance of intracrystalline dislocation mechanisms in D_2 of the northern region is also illustrated in other specimens from the area (46273, 46274, 46275, 46287, 46288).

From south to north, in very similar rock composition, the dominant deformation mechanism on the grain scale for D_2 changes from pressure dissolution to crystal plasticity. The only obvious parameter to change in a similar direction is temperature whose maximum is closely associated in time with D_2. In the north pelitic rocks are schistose in part and contain garnet, albite and chloritoid. To the south, pelites are phyllitic and are almost exclusively quartz-phengite-chlorite rocks with minor tourmaline, zircon and opaques. A lowering of temperature to the south is indicated. Other important variables, such as strain rate, are unknown but the influence of temperature may be critical.

3.6: Minor Structures of the Third Deformation Event

Small scale F_3 structures are extremely rare in pure quartzite and interlaminated quartzite and mica. They are only well developed in micaceous quartzite and are virtually restricted to outcrops of this material in the Wilmots, Starfish and Detached Peak area. The micaceous quartzite is usually a multilayer with varying proportions of mica with some layers being nearly pure quartzite. Wide variations in fold geometry occur depending upon the local sequence. Groupings of different compositions are quite variable as are layer thicknesses and in small outcrops D_3 folds can vary from open to very tight, sub-angular to rounded and moderately flattened to a near class IB profile (Figure 3.17 b). The nature of D_3 folds in micaceous quartzite is very similar to D_2 folds in the same rock type. The complete overlap of styles requires careful study of overprinting and other correlation parameters to distinguish the two events.

Second to third order D_3 folds are known in quartz arenite units but congruous, minor parasitic-folds are very rare. D_3 folds in pure quartzite are open, with rounded hinges and may be associated with a weak generated surface. The major D_3 fold in flaggy quartzite on The Starfish has no parasitic folds except in micaceous horizons.

3.7: D_3 Microfabric

The range of fabrics produced during the third deformation event almost completely overlaps that of D_2. Differentiation along crenulation cleavages is less marked and intensities of development of other fabric types are generally lower. An exception is found in the closure of the major D_3 fold on the Starfish where complete recrystallisation of quartz is induced by this event. The fine dissolution seams, so characteristic of D_2 in the eastern Franklands, also formed in D_3 but on a much reduced frequency (46262, 46246, 46248, Figure 3.25 b). S_3 is also sparsely represented by the mica band morphology (46255).

S_3 reaches its peak development in the flaggy quartzite of the Starfish. Gradations are found within this rock type from a strong D_1 fabric parallel to S_0 (46266, 46276) to a strong D_3 fabric axial planar to the major D_3 fold. S_1 consists of a perfect alignment of tiny phengite flakes which are discontinuous and separated. Quartz grains average 0.2 mm in size and dimensional preferred orientations are variable though generally good; axial ratios are commonly between 2/1 and 3/1. The composition of the flaggy quartzite demonstrates derivation from a very pure sediment but the initial grainsize is unknown as the fabrics are invariably totally recrystallised. Rare undulose old grains up to 0.8 mm are seen but silts or fine sands are thought to be the likely parent sediment for most of the flaggy material to account for the extreme degree of recrystallisation. The intense S_3 fabric in this area is believed to have been superimposed on the typical S_1 fabric described above with 0.2 mm quartz grains and sparse micas. A subgrain development within the S_1 grains is the first microscopic expression of S_3. The subgrains have a weak dimensional preferred orientation in S_3. An intermediate stage between this and total readjustments of the quartz fabric to D_3, is not seen as S_3 intensifies rapidly on the limbs of the F_3 structure. Specimen 46278 represents the next stage where there is no sign of quartz elongation in S_1 but only a slight preferred orientation in S_3 is detected. This structure is best seen in low light conditions in plane polarised light where a distinct alignment of quartz/quartz boundaries is evident. Crystallographic preferred orientation of quartz is weak in S_3. Despite the weak fabric, quartz grain boundaries are sharp and there is no sign of hazy boundaries typical of subgrain/new grain transitions. Mica flakes remain perfectly aligned in S_1 and thin trains of mica rarely exceed 0.5 mm in length, the majority being tiny separate flakes. Several specimens from The Starfish show very strong dimensional preferred orientation of quartz grains in S_3 (46267, 46268, 46270, 46277) and some display marked crystallographic orientation as shown by high percentages of grains showing maximum interference colours in certain orientations (Figure 3.22 b). In all examples of strong S_3 microfabric there are micas parallel to S_1 and in 46270 all micas are parallel to S_1 despite an intense S_3. Bedding is defined in 46268 by thin bands of mica (0.05 mm to 0.1 mm in thickness) which are perfectly aligned in S_1 parallel to S_0. The quartz rich layers between have, parallel to S_3, tiny micas which are much shorter than the quartz grains and, therefore, could not have exercised a marked control on the growth of the quartz grains. 46277 shows a near equal distribution between micas aligned in S_3 and those in S_1 but no evidence of rotation of S_1 to the S_3 direction is noted. Mica films of up to 0.6 mm in length lie in S_3 in association with tiny flakes.

S_3 quartz fabrics appear to be related to deformation of quartz grains recrystallised during D_1. The D_3 strain leads to subgrain formation and probably to subgrain rotation to produce new grains. Intracrystalline dislocation deformation predominates and there is a good correlation between the strength of dimensional and crystallographic preferred orientation. There is no evidence as to the nature of the mechanism responsible for the production of the mica allignment during D_3. Progressive rotations are not seen and it is difficult to correlate the intensity of the quartz fabric with the percentage of mica alignment in S_3.

The D_3 fabrics described above give rise to a strong grain expression in outcrop which often passes directly into a crenulation in adjacent micaceous quartzite layers. This transition is also recognised in several thin sections (e.g. 46269) where the nature of the early fabric and layer composition exert a dominant influence on S_3 cleavage type. In nearly pure quartzite layers, a dimensional preferred orientation of quartz grains is developed. Micaceous layers are crenulated and some differentiation has occurred. S_3 crenulations are associated with minor differentiation and production of a metamorphic layering. In a haematite rich pelitic layer (46261), differentiation has increased the haematite concentration along crenulation cleavage planes (Figure 3.20 b). Microfolding has locally been associated with both differentiation along the limbs of folds and plastic deformation of quartz grains at the fold closures (46263) where small, new grains are elongate parallel to the axial plane. A further D_3 fabric variant, is the rare development of a 'sandstone type' cleavage where mica and quartz beards define S_3 in association with anastomosing dissolution films (46262). The beards may owe their origin to recrystallisation in D_1 and have been modified in D_3.

3.8: Minor Structures of the Fourth Deformation Event

Profiles of D_4 folds vary from open to tight and are rarely isoclinal in quartzite layers in phyllite. Parallel to slightly flattened parallel folds are found in quartzite layers and where interfoliated phyllite layers occur they show characteristics of class 3 folds (Ramsay, 1967). Thus folds are propagated through multilayers of quartzite and phyllite as their total profiles are equivalent to a class II (similar) fold. Typical F_4 profiles are shown in figure 3 of Powell (1969a) and Figure 6(a) of Maclean and Bowen (1971) and Figure 3.23 a of this work. Considerable disharmony occurs in D_4 in interlayered quartzite and phyllite (Figure 3.23 a) and appears to be repeated on several scales (see Powell 1969a, Figure 3e). The degree of disharmony is clearly related to the competence and thickness of quartzite layers and their separation.

The fourth deformation event can be considered the dominant event in the region for it controls the orientation of the important $S_0//S_1$ surface. During D_4 major rotations took place such that S_0 now varies from flat lying to vertical. Because D_4 is the last major deformation, apart from very large scale 'drag folding', the attitude of structures is more constant. Folds of this generation have, for the most part, upright or steeply inclined axial surfaces which strike north/south at the Gordon and gradually change orientation until they lie east/west at the southern end of the Frankland Range. The minor axial surfaces are fanned about major antiforms with wave lengths of the order of one kilometre. Local inhomogeneity can produce axial surface dips of less than 40° particularly in interlayered quartzite and phyllite; in this material single fourth generation cleavage surfaces have been traced from a dip of 30° to vertical within 1.5 metres. F_4 hinges generally plunge at around 20° to the north or west but certain zones show considerable variation in the amount of plunge. Individual hinges have an irregular wave form while lying in the same axial surface and may be explained by variable deformation, in the one event, to produce hinge culminations and depressions. Other examples are produced by the interference of D_4 folds with earlier folds where hinge lines of the two events are slightly oblique. A D_4 fold on the thinned limb of an earlier fold may decrease in amplitude at the thickened closure and change in plunge.

3.9: D_4 Microfabric

The fourth deformation event is widespread and its effects can be seen in the majority of outcrops. As with microfabric of other generations, the intensity of D_4 on the microscopic scale is markedly heterogeneous. Also the range of morphological types is similar to D_2 and D_3 structures and similar processes have operated. D_4 microfolding has affected all rock types from very slightly micaceous quartzite (mica <5%) through to pure mica layers in phyllite. Metamorphic differentiation is not associated with either extreme of this spectrum of folding and is usually found in micaceous quartzite and phyllite. Differentiation in D_4 never occurs on the scale seen in the D_2 crenulations. In rocks of low mica content, deformation of small D_1 new grains and, in highly strained states, some remnant grains (46230) has led to a high degree of dimensional preferred orientation in the axial surface of microfolds (46220, 46221, 46222, 46224, 46229). The process passes through a stage of development of subgrains in the small grains elongated in D_1 (Figure 3.24 a). Presumably the subgrains are rotated sufficiently by the deformation to produce new grains. Several examples are well advanced in their state of recrystallisation, containing distinct new grains with well defined boundaries. Invariably fine, separate, micas remain parallel to S_1. Frequent transitions occur in 2 - 4 cm wavelength folds from marked axial planar alignment of quartz in the closures to an orientation in S_1 on the limbs (46220, 46230) and in well recrystallised specimens local pockets of S_1 fabric may be preserved (Figure 3.24 b, 46229). Lithological control on fabric type is often well illustrated on the thin section scale. Quartzite and mica multilayers alternate from quartz grain fabric to closely spaced crenulations from layer to layer. Thinly laminated portions of quartz rich rocks may show small microfolds and limited differentiation, whereas the dominant fabric is derived by recrystallisation of small quartz grains (46220).

Coarse crenulation cleavages are extremely well developed in micaceous quartzite which often may only contain 5 - 10% mica (46227, 46228, 46234). Mica seams on the limbs of symmetric or asymmetric microfolds are usually spaced at 2 cm intervals and range from 0.1 to 0.5 mm in width with the most common value at 0.2 mm. Crenulation cleavages of this type produce a plane of low cohesion in the field which may penetrate outcrops metres in extent, thus emphasising the weakness imparted by thin mica seams. The mica-seams rarely include quartz but the main evidence for quartz dissolution is the reduced grain size adjacent to the differentiated layers and the production of layering independent of sedimentary features. In lithologies with less than 10% mica, considerable removal of quartz is required to produce the seams. Within the mica layers, the rare quartz grains tend to be very thin almost needle like. A 0.2 mm wide seam in 46228 bifurcates at one point to surround a thin zone of plastic deformation of quartz which is recrystallised on a very fine scale (0.02 mm). Apart from undulosity there is generally little sign of dislocation deformation mechanisms operating in micaceous quartzite. Transitional compositions show mixed behaviour of quartz grain recrystallisation by internal dislocation and diffusive-mass transport. Specimen 46229 shows behaviour ascribed to folding of multilayers by Cosgrove (1976) where, on the limbs of 0.5 cm wavelength folds, the quartz layers are thinned by dissolution thus enhancing the mica layering. Quartz grain alignment is found in the fold closures.

A strong correlation between microfolding and metamorphic differentiation is noted in the micaceous quartzite supporting the proposal that mineral transport depends upon stress gradients created in folds. In several examples, however, differentiation appears to be independent of microfolding though such a process may have initiated the process. 46226 contains about 3% mica and has 0.1 to 0.3 mm thick seams at 1 to 1.5 cm intervals penetrating the hand specimen. Discontinuous seams may occur within the more extensive zones. In detail some of the seams are markedly indented and stylolitic in form, which, combined with the lack of deflection of S_1, suggests a pressure dissolution origin without folding (Nichelsen, 1972). Seams without an associated crenulated cross-lamination are found in rocks which also contain differentiated layers related to microfolds. This latter case may represent the migration of the zone of dissolution beyond the original limb of the microfold thus destroying the evidence for its mode of inception. Specimen 46226 is probably a limited expression of the mica band type of cleavage seen in D_2. The fine seam dissolution features are also found in D_4 in the nearly pure quartzite of the eastern Franklands (Figure 3.25 a, 46246, 46247).

An S_4 crenulation cleavage is almost universal in its occurrence in phyllite outcrops in the region. Intense development penetrates whole exposures with planes of weakness spaced from 0.5 to 2 cm and many finely spaced discontinuous zones. Differentiated mica layers, about 0.5 mm in thickness, are usually associated with the peak of intensity. The normal expression of differentiation is a thin black line in thin section (Figure 3.26 a, b) which nonetheless imparts a pronounced weakness to hand specimens and outcrops. Gradations occur from initial flexing of a strong D_1 fabric (46239) to microfolds with differentiated and attenuated limbs, though with limited concentration of mica (46239, 46235, 46236, 46238). The degree of differentiation is clearly a function of quartz content of the original fabric. If quartz is low then microfolding occurs without extensive production of a layering which is necessary for the formation of a cleavage. Pure mica layers are only folded (Figure 3.26 a) in contrast to slightly quartz-bearing layers which produce a fine dark seam cleavage (Figure 3.26b). With a quartzose phyllite the shape of quartz grains approaching a microfold is strongly modified by dissolution and the cleavage planes are free of quartz (46237).

Some indication of chemical readjustment during crenulation in D_4 is given by Boulter and Raheim (1974). Their S_3 equates with S_4 of the present work and several microfolds and crenulations were examined and in all cases chemical differences could be related to folding. Pure phengite layers generated in D_2 were crumpled in D_4 and departures from S_2 chemistry were related to position in the folds (Boulter and Raheim op. cit., Figure 4a, c and 6b). Stress gradients associated with the folding and changed metamorphic conditions were responsible for movement of species such as potassium, sodium and iron. Most of the crenulated rocks discussed here contain chlorite and phengite, and further evidence of chemical changes induced by microfolding might be obtained by comparing chlorite/ phengite ratios in the microlithons and the fold limbs. Unfortunately chlorite is normally such a minor component that this proved impossible.

3.10: Minor Structures and Microfabric of the Fifth Deformation Event

D_5 folds are very much restricted to the phyllite where very thin quartzite layers, usually less than 2 cm thick, are affected. Many outcrops of this rock type contain D_5 folds though no major folds of this phase are known to exist. Wave lengths of F_5 vary from a few cms to a few metres with the common form surface being

S_4 or $S_0//S_1$. F_5 hinges generally diverge by 20° in a clockwise direction from the F_4 hinges and their axial surfaces dip at moderate to steep angles to the east in the Gordon Dam area. A very fine lineation on S_4 in this area is the common expression of D_5 and pitches steeply to the northeast. D_5 axial surface bearings follow the rotation of the main ridges. A weak crenulation cleavage is produced axial planar to S_5 and is widespread in phyllite.

CHAPTER FOUR

MAJOR FOLD GEOMETRY AND SUB-AREA DESCRIPTIONS

4.1: Nature of Structural Domains

The structural information for the whole of the region has been presented on a series of eleven detailed maps (see Figure 4.1 for key) which do not always equate with boundaries of homogeneous, cylindrical domains with respect to bedding. For convenience of description some structural maps may be grouped together whilst some single map areas may be analysed in two portions. Structural trends rarely change rapidly along the length of the Wilmot and Frankland Ranges and subdivisions within a generally gently curving structure are somewhat arbitrary. Sub-area definition, therefore, often rests upon features such as the occurrence of major, readily recognisable first folds, the presence of an outcrop penetrative, second-generation cleavage or a constant-attitude, enveloping-surface to second order D_2 and D_3 folds. If a distinct grouping of structural attributes can be recognised then this will be used as the basis for defining a sub-area. Such a unit may cover a large arc length of the Ranges and hence in classical terms may form two or more statistically homogeneous macroscopic domains, because of the degree of fold axis variation from one end to the other. In delineating domain boundaries, further considerations may include the effects of size and shape of a sub-area on the presentation of data. The region from The Starfish to The Bell encompasses two homogeneous domains of bedding and though covered by two separate maps is best described as a unit because information from The Bell is critical in determining the generation of the major fold on the Starfish. The eleven structural maps are therefore discussed under nine headings (see below) which take into account the various needs for combination or subdivision in description. Some trend variations may be the result of disharmony coupled with low-angled cross folds. Major D_1 folds are thought to have north/south fold axes in the northern Wilmots but the bedding π diagram shows a swing to east of north brought about by the incoming of micaceous quartzite where D_3 intermediate and small scale folds have different axial trends. Depending upon the extent of the disharmony, such effects may or may not warrant a separate description. Considerable structural variations occur between flaggy quartzite/micaceous quartzite units and the massive quartzite derived from quartz arenite. The former lithologies, however, do not outcrop particularly well and thus require a separate account on only a few occasions.

4.2: Sub-area Descriptions

i) Sub-area 1: Gordon Dam to Serpentine Dam. The region covered by the structural maps, Figures 4.2 and 4.3, was the most intensely surveyed of the whole project area. Ready access, excellent exposure in engineering workings and a range of rock types required a concentrated effort but unfortunately some aspects of this region were atypical. A structural characterisation applied here, had, therefore, to be extended elsewhere with caution. Structural trends continue from the vicinity of Mt. Sprent in a fairly predictable manner following the general swing of statistical fold axes seen along the whole mountain chain (compare Figure 4.4 with Figure 4.9 a). Late cleavages (S_3 and S_4) show trends similar to those of the area immediately to the south, apart from a small strike rotation (compare Figures 4.5 and 4.6 with Figure 4.9). The main departure in the Gordon district is that the majority of fold enveloping surfaces and major lithological units are gently inclined.

Despite structural complexity, the outcrop pattern is fairly simple and any complications are largely the result of the shape of the landforms and the geometric relation of the layering to topographic features. The disposition of the rock units appears to be significantly influenced by a major D_3 fold (1st or 2nd order in present usage) whereas in most other sub-areas this deformation is not so important. From west to east in this area we have several zones distinguished on the basis of the attitude of the layering (see Figure 4.7a and b). At the western margin, massive quartzite dips very steeply to the east and youngs to the east; this passes into a zone of symmetric folds with an enveloping surface dipping west at about 45°. The next zone eastwards consists of flat-lying overturned beds with 'S' shaped, parasitic, D_1 folds and which gradually turns to layering dipping about 35° east, striking north/south. Where younging evidence is available this latter zone is overturned but the analysis is complicated by disharmony as the section is largely in micaceous quartzite. The most eastward unit again dips steeply east and continues trends seen on the lower slopes of Mt. Sprent and in The Bell/Detached Peak area. Both steep zones are narrow in the limits of the Gordon and Serpentine sub-area but they can be demonstrated to be much thicker by traverses down the Gordon River to the west and over Detached Peak to the east. The consistent asymmetry of D_1 folds and younging directions on the flat belt and the gently east dipping section demonstrate that they are both part of an overturned limb of a major D_1 fold. The other limb is the steep, east-younging, western-most portion of the area and the belt of symmetric folds is in the closure. All D_1 minor fold vergences and facings are in agreement with an upward closing, downward facing major D_1 fold. Though incomplete, and affected by major D_3 folding, the overall structure appears to have an axial surface that dips moderately or steeply to the east (50°to 60°approx.). The present attitude of the overturned limb is a result of D_3 and D_4 folding; D_4 brought about a rotation of the whole unit as seen in the Gordon/Serpentine area. Some problems were encountered in the definition of the attitude of the major fold's axis. Examination of π diagrams, constructed from bedding readings in micaceous quartzite units, shows a consistent fold axis bearing of 025° with a variable plunge between 10 and 26° (Figure 4.4 c, d, e from sub-divisions I, II and III). However, the massive, purer quartzite unit gives a π pole at about 002.00 with perhaps a very slight (<5°) northerly plunge (Figure 4.4 b). Apart from these readings, many were taken of bedding in the interlayered quartzite/phyllite units. A combined plot of all bedding readings from the Gordon/Serpentine area shows both the northerly and the 025 trend (Figure 4.4 a). The major D_1 fold axis is north/south whilst D_3 and D_4 folds have a slightly oblique trend. Minor D_1 folds are too few in number to assist with the determination of the fold axis orientation of the major structure. The minor folds have also been disoriented by later events and probably show some degree of rotation towards the D_1 stretching direction. In the flat lying belt the S shaped parasitic folds have thickened common limbs and boudinaged long limbs. A good microscopic fabric of aligned quartz grains (detrital and recrystallised) and mica is found axial planar to folds of this nature on the east thrust face of the Gordon Dam.

D_2 folds only occur as minor structures, having little influence on the outcrop pattern. Within phyllitic units, S_2 is a very distinctive, well-differentiated crenulation-cleavage but it is not universally present and appears to be localised around heterogeneities. F_2 hinges tend to be plunging more north/south than later hinges (Figures 4.5 and 4.6) except in areas of high strain in D_3 and D_4

where F_2 hinges lie closer to F_3 and F_4. Initial variability of hinges is noted which, because of the general gentle inclination of units, gives rise to bearing rather than plunge fluctuation. Complex refolding of F_2 by F_4 and/or F_3 produces irregular hinges where F_2 and F_4 are usually less than 20° apart in bearing. S_2 and the axial surfaces of D_2 folds have been rotated by disharmonic D_3 and D_4 folds and this has led to a large overlap of attitudes of the S_2, S_3 and S_4 crenulations which may all have similar morphologies. Further variation is brought about by refraction from micaceous quartzite to very slightly micaceous quartzite. In the Serpentine Quarry, S_2 refracts from a dip of 25° to 47° in adjacent beds. Another problem in correlation, even over small distances, concerns the pre-D_2 variability in attitude of bedding surfaces around second and third order D_1 folds. S_2 may dip more or less steeply than bedding within small areas, independently of D_2 folding.

S_3 development is quite heterogeneous in this region, being well represented in micaceous quartzite but very weak in phyllitic units. S_3 dips very steeply to the west/northwest but is noticeably affected by pre-existing variability in bedding attitudes. Where bedding is presently gently inclined, S_3 dips about 50 to 60° to WNW and increases in dip over steeply dipping beds. When S_3 transects third order F_1 closures the effect is marked as seen in micaceous quartzite on the ridge between the two dams. In micaceous quartzite, S_3 is a crenulation cleavage with limited differentiation. F_3 hinges commonly plunge moderately on bearings between 020 to 050°.

An analysis of F_4 vergence indicates the presence of several third order D_4 folds in this region in the micaceous quartzite and quartzite/phyllite units. These structures are disharmonic on the major D_1 fold and taken alone cannot define the nature of the major D_4 fold. Extrapolation of information from areas to the south is required to assist with the later interpretation. The most common attitude of S_4 is a dip of about 60° to the east southeast (Figures 4.5 and 4.6) but the attitude ranges from a very steep dip to the west northwest to about 20° dip to the east southeast. Greatest variation occurs in phyllite/quartzite outcrops where gross refraction causes the mean dip to be lower than in micaceous quartzite or quartzite. In single outcrops of quartzite/phyllite, S_4 dip may range from 21° to 71° and refractions of 30° in dip are not uncommon. D_4 structures would be the most commonly recorded features probably because of their frequency of development in phyllite/quartzite outcrops which access roads have tended to cut at low angles to the S_4 strike and F_4 hinge bearing. S_4 usually strikes between 190 and 220 and F_4 hinges plunge at low angles to bearings between 010 and 040. Hinges may vary in plunge within a axial surface to give some dispersion on a projection and D_4 folds do plunge to the north northeast and south southwest. F_4 and F_3 hinges tend to lie clockwise round from F_1 and F_2 but the groupings are not distinct. S_3 and S_4 overlap in orientation because S_3 may be disoriented in D_4. If S_3 is rotated but no S_4 is generated it is possible that S_3 could be labelled S_4 and correlation problems may have occurred this way.

Overprinting of S_2, S_3 and S_4 together is very rare, as is S_2/S_3 and S_3/S_4 interaction. S_4 very commonly overprints S_2 but the above difficulties coupled with the overlap of S_2, S_3 and S_4 crenulation morphologies causes some difficulty in correlation. It is more usual to have a section where bedding and a low angled crenulation consistently dip at low angles and are overprinted in different portions by two variably developed crenulations of quite different attitudes. This is the situation between the Knob Gate and the road junction overlooking the Knob Quarry. Other critical time relationships are seen in the Serpentine Quarry where D_1 boudins have been overfolded in D_2 and then transected by S_3. Within the quarry S_2 and S_3 are both common yet direct overprinting or small scale folding of S_2 are not found. Two groupings of crenulations occur, with some overlap in a zone of constant bedding dip. One cluster strikes between 070 to 120 with dips of 20° to 30° and the other at 020 to 050 with 60° to 70° dips. The former group, when the bedding is rotated to the same attitude as the steep bedding on Mt. Sprent and Detached Peak, lies very close to the S_2 attitude for these areas. The low dip crenulation is, therefore, taken to be S_2 and the steeply west dipping crenulation S_3. Using the π axis from bedding readings near the Serpentine Dam, the S_2 cleavages all fall on a broad cone which relates the 090.20 type attitudes to the 160.50 orientations demonstrating a rotation in D_3.

4.2: ii) <u>Sub-areas 2, 5: Mt. Sprent to Koruna Peak including the Central Wilmots.</u> The most commonly developed structures in the section from Mt. Sprent to Koruna Peak are of the second generation with S_2 being almost universally present in the major rock type of slightly micaceous to pure quartzite. In outcrop, S_2 invariably obscures S_1 and only rarely shows a crenulate nature, presumably as a result of a poor development of platyness during D_1. Many field features, however, point to the post-D_1 timing of the predominant structural surface in the Wilmots. At 39680-72887 and 39670-72865 isoclines in quartzite, several metres in extent, are transected by the cleavage. At 39685-72900 and 39620-72780 tectonic boudins attributed to D_1 are cut at high angles by S_2 such that this cleavage is nearly parallel to the plane of separation of the boudins. Textural evidence further supports the superposition of this surface on an earlier fabric as is particularly well illustrated by 46292 (Figure 4.8).

Here dimensional preferred orientation is found parallel to bedding (S_1) and parallel to S_2. Specimens 46273, 46274 and 46275 show varying degrees of interaction of S_1 and S_2 from near total quartz grain alignment in S_1 (46275) to a high percentage in S_2 (46274) to a near equal relationship (46273). Fine seams of mica generally follow S_2 in all but the purest quartzites and weathering along these gives rise to its prominent field appearance.

Most of the measured S_2 surfaces dip almost directly east at between 25° and 55° (Figure 4.9 c) and the constancy suggests little distortion after D_2 or large scale body rotation. Some minor to intermediate refolding of S_2 has been noted but appears to be limited. It is possible that such occurrences might escape attention because S_2 may be obscured by later structures. Despite the regularity of S_2, the major geometrical features of the Wilmots are somewhat unclear. From 39850-73240 to 39876-72675 the eastern slopes are covered by thick scrub and only widely placed traverses have been made across the axis of the ridge. All indications are that the overall situation involves compositional layering younging and dipping steeply to the east (averaging 60°) (Figures 4.11 and 4.12). The general swing in strike of all structures continues through the length of the area under discussion. An inspection of the topographic map would suggest a trend change at the top of the Koruna sub-area (30605-72605). The π pole diagrams of bedding readings (Figure 4.9 a,b) shows this topographic kink to be unrelated to structure which shows a trend change in the 39696-72968 to 39678-72793 region. For the main spine of the Wilmots the major structures would appear to be non-plunging on a bearing 352. In the vicinity of Mt. Sprent itself a 10° plunge to 004 is shown by bedding readings (Figure 4.9 a). This variation may also be apparent as it could reflect the increasing importance of the slightly oblique structures which become more common on Mt. Sprent. Minor hinge lines and/or intersection lineations of all generations do, however, demonstrate a clockwise rotation of 10° to 20° from the Wilmots to Sprent (Figure 4.10).

Using the π diagrams, an attempt was made to analyse the plunging part of this region with a down plunge profile. 004.10 was used as the projection line and the profile plane attitude was 274.80 (Figure 4.13 d). Field observations suggested a near vertical attitude for the large band of phyllite running north-south immediately west of the summit of Mt. Sprent. On the profile it appeared as a horizontal tongue with steep layering in the quartzites abutting the contact directly. This situation is clearly unrealistic and emphasises the problems of profile drawing in multiply deformed regions. Despite the approximate coaxial nature of the folding events, the π plot gives a poor projection line which must be in error in bearing and may be in error in plunge. A bearing of about 348° would fit the gross situation of steep dips to the east but the plunge is unknown. It is interesting to note that 348 is close to the hinge line bearing from the main Wilmot section emphasising the possibility of the trend change being related to the incoming of disharmonic D_3. It is obvious that total bedding π plots cannot be used alone in determining a profile projection line, even where the girdle is fairly well defined. This approach would lead to quite erroneous and complex results in the Sprent area. More hinge line readings would be desirable as the present data are too sparse to give reliable estimates of the axial orientation relations between the various phases.

Owing to the uncertainties, it was decided to represent the structure in the Sprent-Koruna area by a series of vertical sections perpendicular to the ridge axis (Figure 4.13). The sections show that the average attitude of the layering on the eastern flanks of the Wilmots is a steep dip to the east. Prominent zones of bedding dipping gently to the east do occur and these are interpreted as being the common limbs of Z shaped, second-order D_2 folds (Figure 4.13 a). S_2 maintains a fairly constant attitude through both the steep and gently dipping limbs and despite the lack of clear evidence for the orientation of S_1, the alternative possibility of a large F_1 is discounted. A parasitic D_1 fold occurs on the gentle limb with Z asymmetry and for the large scale fold to be D_1 in age it must be a second order S shape which does not fit the right-way-up younging of the external limbs. The Z parasitic fold is D_1 in age and has been rotated on the common limb of the large D_2 structure.

Figure 4.13 c shows the nature of the large scale, D_2, Z folds which involve overturning of the common limb. The external right-way-up limb dips at about 70° to the east whilst the inverted short limb dips at about 65° to the west. S_2 refraction is apparent because, on the long limb, dips are around 30° to 40° whilst, on the common limb, values are between 40° and 55°. Limbs are straight and the closure is somewhat angular.

S_3 along the Wilmots maintains a regular attitude and dips steeply to the west, varying in strike from 340 to 020. Local folding in D_4 is noted but overall a bulk rotation during this later event is indicated. Towards the summit of Mt. Sprent, D_3 has produced second to third order folds and S_3 becomes an intensely developed, pervasive crenulation. The large folds are open and angular with steep axial surface dips to the west (Figure 4.13 b). Their exact profile shape is difficult to determine because they occur in association with D_2 folds.

D_4 folds only occur on a minor scale but are invariably S shaped, climbing to the west, indicating the closure of the major fold is in that direction. S_4 is usually dipping steeply east but is refracted in phyllite and micaceous quartzite horizons and disoriented by S_5. The fifth cleavage is entirely restricted to phyllite and dips gently east.

4.2: iii) <u>Sub-areas 3, 4: The Bell, Detached Peak and The Starfish.</u> The Bell/Detached Peak area is essentially a uniformly dipping (175.75) western limb of a third generation large scale synform. Three mainly competent rock units, a micaceous quartzite band sandwiched between two flaggy quartzite units, are bounded to the west and east by phyllite and quartzose phyllite. These units have been mapped across Wilmot Bay to The Starfish where the closure and part of the eastern limb of the major fold is found. When poles to compositional layers are plotted separately for both areas a 20° change of overall trends is noted (Figure 4.14 a and b) as well as the complexity of the distribution for The Starfish. In order to determine the statistical hinge line in this latter zone (for down plunge profile construction), compositional layer readings from immediately around the main closure were plotted on a stereonet (Figure 4.14 c). A conical pattern is indicated for the actual closure but the whole fold very rapidly becomes near cylindrical and there is a lessening of the plunge of the hinge to the north (Figure 4.14 d). At The Starfish, the form of the major fold is most clearly defined by the lowermost flaggy quartzite horizon because it is least affected by second phase folding. All compositional layer data from this rock type would indicate a statistical hinge at 43.353. However, leaving out data from the conical closure and considering the northwards decrease of plunge, a profile was constructed about an average fold axis for The Starfish of 30.350. A total plot of compositional layers from The Bell shows a hinge at 20.007 (Figure 4.14 b) and on this basis, also using the continuity of rock units across Wilmot Bay, a profile of The Bell/Detached Peak was constructed and added to that of The Starfish (Figures 4.15 and 4.16). Such a procedure gives the impression of a very long western limb to the major fold whereas the plunge may have flattened out and perhaps these two profiles should be superimposed one behind the other at approximately the same level. The difficulty arises because the hinge for The Bell is derived almost entirely from areas with common second folds and these may well be plunging at a low angle on the planar limb of a non-plunging third generation fold. The Bell profile in effect is a profile for the minor second generation folds of that area.

Only minor D_1 folds are noted in sub-area 3 and hinge lines and axial surfaces are disoriented by later events (Figure 4.14 e). An interesting feature of the F_1 hinge distribution is the scatter on the regular, western limb of the major D_3 fold. Presumably variable stretching during D_1 caused different amounts of rotation of the minor hinges. The most consistent surface of both sub-areas 3 and 4, in terms of orientation, dips 40° to 65° to the west and commonly strikes 350 to 020 (Figures 4.17 and 4.18). It is often seen to crenulate S_2 but in mica poor rocks again takes on a penetrative appearance with little or no trace of an earlier tectonic surface. The attitude of S_3 (Figure 4.14 g) fits the axial surface direction of the major fold and folds associated with this surface have congruous asymmetries with the larger structure (Figure 4.16). Again there is a very marked spatial relationship between the intensity of cleavage development and the closeness to the closure of important folds. On the long ridge between Sprent Basin and The Bell, S_3 is usually a very weak crenulation but following this limb across Wilmot Bay the surface becomes more noticeable and the intensity suddenly increases at the closure where it becomes penetrative. Such variations probably explain why S_3 rarely is found to fold or cut S_2 in the micaceous quartzite and uppermost flaggy unit of The Bell/Detached Peak.

The next structural event is distinct in both orientation and overprinting relations with earlier events, yet everywhere it is weakly expressed. In the competent units, S_4 is steeply inclined trending approximately north/south (Figure 4.14 1) and is equated with a moderately east-inclined crenulation in phyllitic rocks. This trend and degree of refraction is very similar to the dominant folds of the Gordon Road to Knob Damsite section and appears to have had little effect on The Bell/Starfish complex of pre-existing folds on an outcrop scale.

The spatial segregation of deformation phases is more apparent here than in most other parts of the Frankland and Wilmot Ranges. D_2 folds are well developed at The Bell but are virtually absent at The Starfish where F_3 is found on a large scale with common parasitic folds. Some difficulties therefore arise in terms of structural correlation even in such a small area. A weak S_3 overprint of S_2 has to be extrapolated on orientation grounds to a zone of strong D_3 structures lacking much evidence for D_2. In a similar fashion the upright S_4 is rarely seen to overprint S_2 but in The Starfish area occasionally cuts S_3 and is hence placed in the structural sequence. Outcrops preserving S_1, S_2 and S_3 are rare and critical for the interpretation of superposition.

4.2: iv) Sub-area 6: Southern Wilmots. The Southern Wilmots covers a wide range of structural domains from the extensive outcrops of quartzose phyllite in the old Serpentine Valley, through a zone of steep, eastward dipping, quartzite that extends along the entire eastern flanks of the Wilmots to an area of generally moderate west dips on the crest of the range (Figure 4.19). The π diagram for the sub-area (Figure 4.20 a and b) indicates a very gentle plunge for the gross structure, but because of the spread in the girdle an accurate determination of the fold axis is impossible. The contoured bedding readings suggest a value at about 154.05 for the axis. An analysis of the minor fold hinges (Figure 4.20 b,c,d,e,f) for the various deformation events, suggests that the folds plunge gently on either side of horizontal and the net effect of these undulations could be essentially a non-plunging major structure. Because there are significant structural differences between the west and east dipping portion of this area (Figures 4.19 and 4.21) they will initially be described separately. Their boundary is the axial surface of a first order, fourth generation fold and all earlier structures are rotated from one area to the other. Structural sequences have to be established in each portion and then compared. A complicating factor is that the west dipping zone is composed almost entirely of pure, massive quartzite whereas the eastern sector contains a large proportion of micaceous quartzite and phyllitic material. Moderate crenulations may be readily developed to the east yet not at all in the competent pure quartzite to the west.

Eastern zone: The eastern part of the Southern Wilmots section (Figure 4.21 a) may be superimposed on the section near Koruna Peak (Figure 4.13 c) as they are both directly along strike from one another in an essentially non-plunging structure. This relationship is sketched in Figure 4.21 c, demonstrating the opposite sense of vergence of the second or third order D_2 and D_3 folds. S_2 dips moderately to steeply west. Unfortunately, S_2 is not well developed in the steeply east-dipping quartzite of the Southern Wilmots. The extrapolation along strike, has, therefore, to be made from Koruna Peak to clearly illustrate the D_2/D_3 interaction. Such a procedure is reasonable considering the continuity of dip and strike of the compositional layering which is quite regular from Tombstone Hill to near Mt. Sprent. Specimen 46282 shows that the surface with orientation 335.48 (at 40154-71925) is the third fabric in the rocks, post-dating S_1 and a widely spaced set of dissolution seams (S_2). S_4 is fairly constant in attitude in this zone, apparently transecting all earlier features and is steeper than the layering in the quartzite/micaceous quartzite rock types. In the quartzose phyllite of the valley, S_4 may be dipping at angles as low as 61°; presumably a refraction effect.

Western zone: Here dips are generally of the order of 35° to the west southwest and the most regular feature is a near vertical cleavage that post-dates three different style groups of folds. Despite the problems of correlation based on style, it is thought that two of these groups are sufficiently distinctive to allow comparison with D_1 and D_2 folds in nearby sub-areas where over-printing evidence is available. The D_1 folds are coupled, asymmetric, moderately flattened, isoclines with a fabric parallel to the external limbs. The D_2 folds are the characteristic step like, kink folds with the predominant Z vergence found in most sub-areas. Locally folds with an S vergence, and axial surface cleavages dipping between 50° and 60° east appear to pre-date the main cleavage and are thus assigned to the third deformation. This whole zone can, therefore, either be interpreted as the common limb of a Z verging, second-order, fourth-generation fold or part of the closure of the first order D_4 fold (see section on major structures). Second generation cleavage cannot be identified in this area and S_3 is only rarely developed in association with S vergent folds of a different nature to the small symmetric folds that post date S_4. Patchy development of S_5 and F_5 disorients S_4 but only occurs on a scale of centimetres. S_5 and S_3 are, however, virtually coplanar and in several instances cannot be distinguished on overprinting grounds. In such situations the characteristics of the associated folding are used for differentiation.

One particular feature of the Southern Wilmots points to it being the location of a major D_4 fold; this is the intensity of the near vertical S_4 cleavage, which dips between 75° to the east and 85° to the west in the quartzite and micaceous quartzite of the main ridge. On various scales the near upright folds show a stronger development of the generated surface towards the closure. In many examples the cleavage may disappear on the limbs. The westernmost rocks of the east dipping zone all carry this intensely developed fabric, generally as an outcrop penetrating crenulation. Fabrics in the massive quartzite of the west dipping zone are dominated by S_4 (46220, -221, -222, -223, -224, -225). 46221 has a well defined D_1 fabric of fine micas and elongate old grains (ratios 7 to 1); the latter being remnants of detrital grains divided into zones of recrystallisation by new grain growth along deformation bands. The new grains of D_1 are now all markedly elongate in S_4 at high angles to the length of the old grains. Rare examples are also noted of totally recrystallised complete detrital grains, as defined by faint haematite lines, being elongate in S_1, with all the new grains showing a strong dimensional preferred orientation in S_4. In 46222 parts of old grains are more common but here they show a tendency to be aligned in both directions. This specimen shows perfect alignment of tiny mica flakes only in S_1 despite the strong preferred orientation of small quartz grains in S_4. 46220 and 46224 both show small D_4 folds of S_1 with varying proportions of new grain alignment in S_4. In 46220, the S_1 fabric has been affected by, low angle, weak, pressure solution seams which are pre D_4. Micro-textural evidence, therefore, shows that the predominant fabric is at least D_3. Arguments outlined above, together with a consideration of possible rotations of Figure 4.21 b, indicate that it is D_4 in age.

The Southern Wilmots are unusual in the degree of exposure of the phyllite and quartzose phyllite in the valley floor. A low easterly trending ridge gives an opportunity for the examination of the incompetent rock types. In this material, S_2 is a very variably developed surface which, whenever it appears, involves marked metamorphic differentiation in association with a crenulation cleavage (46281, 46283 to 46286). Overprinting relationships show that this type of surface is always the first crenulation cleavage to form; the later crenulations are only rarely differentiated to a similar extent. S_3 and S_4 in the quartzose phyllite maintain their usual orientation relation with S_4 at a lower angle than normal in some localities.

4.2: v) Sub-area 7: Tribulation Ridge. This sub-area includes a section of ridge which trends at right angles to the backbone of the mountain range. The main ridge segment seen here is predominantly pure, cross-bedded, quartzite with intercalations of slightly-micaceous quartzite. Immediately to the west is a 500 metre thick unit of phyllite which occupies a col separating the two ridges. The phyllite is followed by a micaceous quartzite, pure quartzite, micaceous quartzite, phyllite, quartzite sequence in a south-westerly traverse along the subsidiary ridge (Figure 4.22). Throughout the sub-area compositional layer dips are generally very steep, often between 80° and 90° to the north-east (Figure 4.23 a). The thick units on the minor ridge also appear to dip very steeply in the same direction. In the northern section of the Coronation Peak-Double Peak map (Figure 4.24), the general layering dips very steeply to the southwest. The Tribulation area is a continuation of this trend where the layering reached the vertical and then became structurally overturned. As this sector is near the closure of a major, tight D_4 fold which has an axial surface dip to the northeast, a west limb, dipping steeply east, is to be expected. Despite east dips, minor D_4 folds also verge east locating the major antiformal closure still further to the east.

Minor structures of the fourth deformation event are most obvious in the area. S_1 is well developed in thin section but is usually only recognised in the field in microlithons of coarsely crenulated micaceous quartzite. D_2 structures occur in micaceous quartzite and phyllite on the south west trending ridge. A prominent mineral lineation is related to this event and rootless, intrafolial folds of thin pure quartz layers are D_2 in age. S_2 crenulation cleavage occurs sporadically in the phyllite. An S_3 cleavage was only detected at one locality but the high strain in D_4 has meant considerable convergence of planar structures. It is possible that several examples of S_2 and S_3 could have been classified as S_4 because of their low angular relations in this region.

S_4 is intensely developed on the main ridge but its intensity dies off westward. The fourth generation surface is an outcrop penetrating crenulation in slightly micaceous quartzite or a pervasive 'grain' cleavage in pure quartzite. Besides extensive cleavage formation, high D_4 strain is indicated by deformed cross-bedding and minor fold hinges. At the closures of minor D_4 folds, foreset beds are nearly parallel to the axial surface as a result of high strain.

Minor fold hinges within phyllite units tend to be steeply plunging within the S_4 cleavage and have probably rotated towards the maximum elongation direction during D_4. Similar rotation appears to have affected D_2 lineations and minor hinges which also are steeply plunging. S_4 dips very steeply to the north-east with an average value of about 80° (Figure 4.23 c). Some of the vertical and steep dips to the southwest may be of earlier surfaces thus erroneously increasing the value of the average S_4 dip. Lower dips of S_4 to the northeast reflect rotation during D_5 or kinking events.

As noted elsewhere, zones of intense S_4 correlate with increased development of D_5 minor structures. S_4 on the main ridge, is commonly affected by small ($< 1\text{m}\lambda$) open folds with axial surfaces that dip moderately to the northeast. A weak to moderate crenulation, S_5, is best seen in micaceous quartzite axial planar to F_5 (Figure 4.23 d). F_5 hinges follow approximately the same trend as F_4.

4.2: vi) Sub-area 8: Coronation Peak-Double Peak. The Coronation Peak, Double Peak and Tombstone Hill sub-area encompasses the major structural change from steep west to steep east dips of the compositional layering (bedding with S_1 usually parallel). As such, the sub-area will be considered in two portions; Coronation Peak to Double Peak, and Tombstone Hill. The first of these areas has largely very steep dips to the southwest with a flatter zone along the ridges east of Double Peak (Figures 4.24 and 4.25). A stereographic projection of bedding poles (Figure 4.26 a) shows an extensive girdle with a clearly defined statistical fold axis of 15° to 304° which was was used as the construction line for the down plunge profile (Figure 4.27).

This section has been combined with the Tombstone Hill and Redtop to Cleft Peak projections in order to illustrate the nature of the major structures (Figure 4.25).

Sedimentary cross lamination is well preserved in the quartz arenite of the Coronation to Double Peak region but was little used in the determination of facing because of the frequency of first tectonic folds. D_1 folds of the well flattened non-vergent, intrafolial type are common along with many examples of asymmetric, D_1 folds. In the early stages of mapping it was felt that zones of regular attitude, on the mesocopic scale, would have to show consistent facing in order to demonstrate the correct identification of sedimentary cross lamination. Folds can mimic cross lamination completely in morphology and hence great care was taken initially. Prior to this study such features were exceedingly rare in the Tasmanian metamorphosed Precambrian and hence only exceptionally clear examples were utilised. As mapping proceeded, it became clear that sedimentary structures and textures were often preserved. Virtually all the asymmetric D_1 folds are S vergent (looking northwest) which, coupled with the limited younging evidence, indicates the bulk of this zone is on the overturned limb of a major D_1 fold of nappe dimensions.

Characteristic D_2 folds are found to the south of Double Peak in perhaps their best development. Here the widest range within the general kink grouping is seen and they also refold F_1 structures (40185-71530, Figure 3.16 f) to provide unequivocal overprinting relations. This same locality is close to Redtop Peak where S_3 is common and is a clear example of a small second order, third generation fold, folding the D_2 kinks. Figure 4.28 a shows gently dipping quartzite in the foreground with Z F_2 kinks, folds of the same asymmetry occur on the vertical limb of the D_3 fold in the background. The F_3 axial surface dips at about 50° to the southwest. S_2 appears not to be developed at this locality, though S_3 is a sporadic outcrop feature.

Some doubt exists as to the position in the structural sequence of the moderate to steep east dipping cleavage between 40322-71650 and 40283-71614. This is on the more gently inclined portion of the profile (Figure 4.27) which can be interpreted two ways: it is either the common limb of a Z vergent D_4 fold or an open angular D_2 fold. The uncertain surface would be S_3 in the first interpretation or S_2 in the latter. Field evidence of overprinting does not clarify the position of this surface, except to suggest tentatively that it is more likely to be S_3.

Throughout the Coronation Peak-Double Peak zone, S_4 is an intense to well developed grain fabric which can obscure earlier features or hinder their clear recognition. In this high strain region for D_4, the generated cleavage is nearly vertical (Figure 4.26 d), dipping steeply both to the northeast or southwest. From the interpretation of major structures, overall layering dips steeply on either side of a first order D_4 antiform and a near upright fold is to be expected with a steep axial surface dip to the northeast. Away from the closure lower dips of S_4 must be a result of fanning, refraction and later tectonic events. Also, as appears to be the case elsewhere, intense S_4 development has localised the formation of S_5 which appears to require a planar fabric for its initiation. S_5 dips at moderate angles to the northeast (Figure 4.26 e).

Minor fold hinges are approximately coaxial for the important deformation events, though individual hinges may be somewhat variable.

F_1 hinges fall in the same area on the stereoplot as do the F_4 hinges. F_5 minor hinges tend to be a little steeper but have the same bearing as the earlier phases.

In the phyllite unit between Double Peak and Coronation Peak, S_4 and S_5 cleavages are predominant. It is also in this horizon that folded quartzite layers show greatest variability in hinge line plunges. Figure 4.28 b shows a bulge of quartzite within the phyllite as a result of a curved F_1 hinge coming out of and re-entering the hillside.

4.2: vii) Sub-area 8: Tombstone Hill. The major unit of the Tombstone Hill district is a massive quartzite succession dipping 70° to the northeast and which, on the basis of sparse evidence, faces in the same direction (Figures 4.24, 4.25 and 4.27). This unit is at the structural base of a simply layered structure involving massive quartzite - platy quartzite - massive and platy quartzite - micaceous quartzite. Within this sub-area, these horizons form reasonably distinct mappable units but their continuation is uncertain. Groupings of several of these are required for extensions of boundaries to be made to other areas. Deviations from the regularly dipping layered slab are found in the southeast portion of Tombstone Hill where layer attitudes of about 313.50 are common. When all compositional layers are plotted on a π diagram a composite effect is seen. This latter zone of moderate southwest dips, together with other readings in the same unit of massive and platy quartzite, give rise to a great circle distribution of poles (Figure 4.29 a, b) whose π pole plunges 05 to 314. The other aspect of this composite plot fits a conical distribution fairly well with a cone axis plunging 60 to 304, the beds defining an apical angle of 80°. A possible interpretation of the pattern is that it results from the interference of two cylindrical folding events whose axes trend in the same direction but plunge by differing amounts. Either interpretation suggests the interference of shallow and steeply plunging events along approximately the same bearing. When the π plot (Figure 4.29 a) is subdivided, such that readings from the more competent units can be distinguished, it is seen that both trends are equally well represented. This situation cannot be explained in terms of non-coaxial events only affecting the incompetent lithologies whilst leaving the major structure essentially simple in geometry. Because of the composite π plot and the obvious effect of both trends on large competent units, there are difficulties in presenting the structural data of the area on either a vertical section or a profile. Both will distort critical angular relations and neither can give a 'true' representation of the large and small scale structures. Because sections or profiles are useful for visualisation of structures, a profile (60.209) was constructed about a construction line 30.299. This is a compromise between the steep and shallow trends and must be considered only as a means of presenting all the data together in an appreciable form. By taking average values for the two most regularly developed planar zones, viz. 313.50 and 104.70, a β plot gives a fold axis of 292.23 and two possible axial surfaces (as bisectors of the representative limbs), 046.25 and 296.80. Figure 4.27 would indicate 046.25 to be the most likely choice for the axial surface of the fold seen on the lower portion of the profile. This fold is post D_1 and S_2/S_3 relations suggest it is best ascribed to D_2 with some modification in D_3 (and possibly D_4?). The analysis of hinge lines does not help to distinguish which deformation is responsible for a particular trend. Hinge lines of minor folds for each event show quite a scatter of orientations.

S_3 is the most consistently oriented structural surface in the Tombstone Hill area (Figure 4.29 e), despite its variation from intense to almost no development. In pure quartzite it is a grain type cleavage (46261, 46262) or absent, whilst in micaceous quartzite shows a crenulate form (46263). S_3 dips 60° to 70° to the southwest and is usually readily distinguished from S_2 using overprinting relationships. The cleavage produced in D_2 is only well developed in micaceous quartzite where it is found at a low angle to S_0 as an outcrop penetrating crenulation (46259, 4620, 46264). In slightly micaceous rocks S_2 is a series of well spaced (approx. 1 cm) dissolution zones involving considerable concentration of mica (46262). S_2 has been disoriented during later events and has a tendency to form a girdle whose pole is at 313.13. There are some instances where S_2 and S_3 are virtually coplanar and identifications have to be made, via style and nature of associated folds, from outcrops with overprinting evidence. The fourth structural surface is only moderately to weakly developed generally dipping steeply northeast.

4.2: viii) Sub-area 9: Redtop Peak to Cleft Peak. This large sub-area (Figure 4.30) shows a fairly regular statistical fold axis for the region containing mainly west dips, i.e. the ridge running from Redtop Peak to Cleft Peak. Poles to bedding (Figure 4.31 a) indicate a profile construction line of 10 to 289 and a profile plane dip and strike of 80.199. Cleft Peak, however, when analysed separately shows a clear girdle with a π axis of 20 to 287 (Figure 4.31 b) and appears to stand apart when plotted with the rest of the data. Most of the difference is removed by omitting data from the second order D_3 fold on the lower slopes of the ridge trending northeast from Cleft Peak. To illustrate the variation, the Cleft

Peak area has been plotted with a profile of 80.199 on the diagram of pitches of structural surfaces (Figure 4.32) whilst on the interpretation of large scale features (Figure 4.25) this portion is on a plane 70.197. In this latter presentation, the common limb of the third generation fold is extended and appears quite disharmonic. The statistical fold axis for the remote Peak district is approximately 25 to 318 (Figure 4.33) but the definition of this direction is restricted by the lack of variation in bedding attitudes. The profile for Remote Peak has been plotted so as to represent the whole subarea and also to show how it is linked to the Double Peak-Coronation Peak-Tombstone Hill sub-area number eight. The actual profile is, therefore, a composite, constructed about several projection lines and fitted together by keying in reference points.

Gross features that emerge from the profile include the upper portion of a near upright D_4 synformal fold between Remote Peak and The Citadel. Here this fold is open but appears to tighten towards the north. To the north of The Citadel and west of The Lion is a relatively flat belt which is the common limb of a second order D_4 fold verging to the east. Below this point in the profile, compositional layer dips are commonly of the order of 40°to 50°; above, dips steepen progressively to 70° at the Cupola and are near vertical around Double Peak. Younging evidence, together with the asymmetry of D_1 folds, show that most of the west dipping belt is upside down. As this inverted portion extends over several kilometres it must represent the overturned limb of a nappe.

Within the Redtop Peak sub-area there is much evidence for the first deformation event. Cleavages are commonly seen parallel or sub-parallel to bedding which is frequently boudinaged. Coupled isoclinal folds are common whilst larger, non-vergent isoclines within the layers are noted from time to time (see Figure 4.32). Only a small zone east of Cleft Peak shows S_1 at an angle to bedding.

Although D_2 and D_3 features occur sporadically throughout the sub-area (Figure 4.30), they are only commonly developed between the Cupola and Redtop Peak. D_2 structures are also found in many phyllite exposures but are difficult to measure because of their small size and strong overprinting by D_4. Near Redtop Peak, S_2 dips moderately to steeply east whilst S_3 dips gently to steeply west. Bedding maintains a fairly constant 70° dip to the west except on the southeast side of The Cupola where a third order D_3 fold occurs. Within the Redtop Peak/Cupola zone S_2 and S_3 both have similar field appearances (46231, 46232) which in turn is very like S_4, the orientation of which they both approach locally. Problems of identification and separation of these three surfaces occur but fortunately overprinting evidence is sufficient for their determination in most examples. Another problem results from the very variable intensity of cleavages; S_3 may be well developed in part of one outcrop whilst a few metres away no associated fabric is visible (46233). However, S_3 transects several D_2 folds which in turn locally refold F_1 and by this means in a small area, such as from Redtop Peak to The Cupola, a consistent structural sequence can be confirmed.

D_4 structures are the most frequently observed tectonic elements, being found in almost every outcrop. The attitude of S_4 is steep, dipping either side of vertical, and striking west northwest with predominance of east northeast dips (Figure 4.31 e). Variation in the dip of S_4 is caused by refraction and fanning around folds of all magnitudes and also by refolding in D_5. Minor F_4 hinges plunge gently to moderately to either side of the horizontal within the mean D_4 axial surface (Figure 4.31 e). S_4 and S_5 are the most commonly recorded surfaces in the phyllite outcrops. S_1 parallel to S_0, and S_2 are frequently seen but are often too variable in attitude to measure. S_5 is only present where S_4 is an intense planar fabric and is, therefore, virtually restricted to micaceous quartzite and phyllite lithologies. The fifth generated surface dips east northeast at between 8° and 62°, with a clustering at 40°, and shares a similar strike to S_4. Overprinting evidence illustrates that many of the S_5 surfaces with high dips are correctly placed in the structural sequence but there may be some which are irresolvable because of overlap of attitude and morphology. In the latter examples, extension, from nearby exposures with better control, is used to assign surfaces to events.

A smaller second order fold on the lower slopes of Murphys Bluff provides evidence for the location of a second order isoclinal D_1 fold. The area involves a cliff face with a sharp break of slope into the valley; above the cliff is a flat bench. At the lower break of slope layering in quartzite dips steeply east and is right way up. First generation folds have Z asymmetry. On the flat bench, layering dips at about 35° to the west, is upside down and contains first folds of S asymmetry. S_4 transects the whole structure, dipping on average 60° to the east northeast. The best explanation is that an F_1 axial surface runs parallel to the cliff edge, but standing out from it. Near the upper break of slope, the axial surface is folded into the hillside such that in a traverse up the ridge, the axial surface is crossed.

In the Remote Peak area, D_1 and D_4 mesostructures predominate on the east dipping limb of the major D_4 synform (Figure 4.25 and 4.32). Because the enveloping surface dips at about 45° to the north northeast, it is predicted that S_3 is likely to have a steep south southwest dip. A post S_1 grain type cleavage in the quartzite generally dips steeply north northeast but a morphologically identical surface dips steeply south southeast. Considering fanning about the major D_4 fold, this cleavage could be S_4 alone, but the possibility exists that it could in part be S_3. At 40347-71204, a grain cleavage (S_3) with attitude 322.84 is folded by a later event whose axial surface (109.70) is very similar in orientation to the general S_4 attitude.

4.2: ix) Sub-areas 10,11: Greycap, Frankland Saddle, Frankland Peak, Secheron Peak, Terminal Peak. Virtually all the outcrop in these two areas is quartzite with very few occurrences of micaceous quartzite or phyllite. Though the area covered by these maps includes a large swing in trends of statistical fold axes, they shall be described together as the region is unified by the widespread clear definition of major D_1 folds. Accordingly the profile (Figures 4.34, 4.35) encompasses the area of the two detailed structural maps (Figures 4.36, 4.37). The profile was constructed in three parts and joined using keyed outcrops. Between the east face of Secheron Peak to the end of the Frankland Range, 263.20 was used as the construction line, from Secheron Peak to Frankland Saddle, 274.16, and from the Saddle to Greycap, 310.20. To illustrate the change in trends several smaller zones were plotted separately (Figure 4.38 sheets 1 and 2) and the results presented on a sketch map (Figure 4.39). These subdivisions are coherent units and in total do not represent all the bedding readings from the two structural maps (Figures 4.36, 4.37). In general the statistical fold axis is plotted on Figure 4.39 and in some examples may not be the most reliable construction line. The zone 2C has a fold axis of 250.30 as the pole to the best fit girdle of bedding poles, but the minor fold hinge lines and S_0/S_1 intersection lineations plot nearer 260.22. Zone 1 D + E is from a major, close, chevron fold and the concentrations of poles representing the two limbs are not well separated. The fold axis in such a case is poorly defined (Ramsay, 1964) and the value 267.20 is open to question. Minor hinge lines and calculated intersection lineations (S_0/S_1) tend to plot closer to 275 which is in better agreement with information from areas 1 F and 2 A + B. On the scale of the Frankland Range the trend variation at Frankland Saddle is very rapid and occurs in approximately one kilometre. It is difficult to be more precise though the change may take place over a few tens of metres. The total plot of bedding poles from the Saddle to Terminal Peak (Figure 4.40 a) contains no poles to steeply dipping beds that could belong to the north-westerly trend. Figure 4.38, sheet 1 C, shows bedding poles from 200 m northwest of the Saddle to 1000 m southeast, and indicates that the change of trend must be at the Saddle or within a few hundred metres to the northwest.

When the complex structural situation is considered, it is remarkable that the plot of all bedding poles east of the Saddle shows such a small degree of dispersion. Most of the spread of the girdle is derived from the variations of first fold trends as seen in Figure 4.39. The total plot is largely the interaction of major D_1 and D_4 folds, though the latter were only sampled in the western portion of the area which is to the south of Frankland Peak. The small sample of the north dipping limb of the major D_4 and overall isoclinal nature of the first folds, means that one half of the girdle is poorly represented (Figure 4.40 b). A further bias to the sampling comes from the inclusion of readings from the three intensely surveyed sites for palaeocurrent work, which all fall in the zone of close D_1 folds. West of the Saddle again samples are across major D_1 and D_4 folds, but in this area the simple distribution of bedding poles (Figure 4.38, Sheet 1 C) is largely attributable to the D_4 fold. The outcrop and size of the D_1 fold means that this structure does not contribute heavily to the spread of points. The π diagrams are therefore, fairly regular because one event is usually better represented than any other but also the phases must be close to coaxial. This latter point can be fairly readily demonstrated to the west of the Saddle. The girdle and π pole (Figure 4.38, Sheet 1 C) came largely from the D_4 fold but nearly coincide with the S_0/S_1 intersection lineations and the π pole for the immediate area of the D_1 fold on the ridge leading down from Greycap (Figure 4.38, Sheet 1 A). The near coaxiality appears to continue to the east of the saddle. The intervening events D_2 and D_3 seem very poorly developed and are not responsible for much variation in the diagrams.

The major fourth generation fold seen on the profile (Figures 4.34, 4.35) and on the structural maps is virtually upright with a slight inclination to the north or north northeast. There is some evidence for fanning of the fourth cleavage (Figure 4.40 f) about this structure. On the north-dipping limb, S_4 dips to the south or south southwest, but not invariably whilst its predominant dip elsewhere is in the opposite sense. Minor D_4 folds are only found near the closure of the major fold and have the appropriate vergence on the two limbs. S_4 is most commonly a crenulation cleavage but may take on a grain appearance in some outcrops. It reaches a maximum of development in the valley south of Greycap which is along the closure of the major fold. Locally it is common where a very platy S_1 fabric is present. The attitude of S_4 lends further support to the demonstration of rotation of the range. Figure 4.40 f shows two distinct groupings of poles to S_4, the cluster nearest the south mark is largely from east of the Saddle and has a mean strike of 105. The other cluster, from west of the Saddle, has a mean strike of 140 and the rotation closely matches the 36° recorded by the π poles. The most abrupt change is across the Saddle (see Figure 4.36) and points to the localisation of the change in a very small area.

Features belonging to D_2 and D_3 are so sparsely represented that no interaction between them is known (except for small areas of micaceous quartzite) and essentially there is a composite group of structures that can be shown to post-date S_1 and pre-date S_4. Assignment of such structures to a particular event is on the basis of the asymmetry of the folds, the attitude of the generated surface with respect to the enveloping surface and extrapolation of information from the rare micaceous quartzite zones. Along the northwest trending ridge between the Saddle and Greycap there are two groups of folds with differing styles and senses of asymmetry within regularly dipping, often boudinaged quartzite layers. Both sets are overprinted by D_4. The group believed to be D_2 folds are kink like with Z asymmetries (Figure 4.41 a) but with little sign of a generated cleavage. The D_3 group has S asymmetry and individual folds are close to tight and sub-rounded. A crenulation is axial planar to these folds.

S_2 was recognised in the field infrequently. In the rare patches of micaceous quartzite it is an outcrop penetrating, widely spaced crenulation cleavage. In thin section S_2 is characterised by fine dissolution zones in the purer quartzite (46250 to 46252, 46246 to 46248, 46256, 46294, 44370). This morphology is similar to some expressions of S_3 (46255, 46256) and S_4 and it is the normal situation to find two different trends of dissolution surfaces in thin sections. Time relations between the two are difficult to determine and orientation with respect to the enveloping surface has to be employed. In quartzite coarser than fine sand in grain size, any one generation anastomoses around the anisotropy given by the quartz grains and tends to blur the definition of the overall trend.

The major interest in the Greycap to Terminal Peak area is the presence of well exposed, major, first generation folds which also show a wide range of styles. Two style groups exist and generally they are found on either side of a distinct tectonic boundary. It is believed that this boundary is a thrust or metamorphic slide which was active during D_1 and where movement was taken up along a direction parallel to the bedding. The discontinuity is defined closely at 41170-70960 and 41230-70935 and can be followed along this trend to Frankland Saddle but cannot be traced east of Frankland Peak. Above the slide, S_1 is parallel to S_0 except in the immediate region of a large fold closure and boudinaged layers are common (Figure 4.41 b). First folds here are rounded in the closure and isoclinal (Figure 4.42). Below the slide, S_1 is usually at an angle to S_0 and, though most D_1 folds are isoclinal, they are not rounded in the closure (see profile, Figure 4.35). The change from $S_1 \wedge S_0$ to $S_1//S_0$ is very sharp being defined in the field to a zone less than 0.5 m thick. Strain during D_1 intensifies towards the slide as seen in the field expression of cleavage and on the basis of the analysis carried out in Chapter Seven. Recrystallisation in the Franklands reaches a peak in this zone for pure quartz arenite of apparently normal initial grain size.

The contrasts across the zone may suggest that different deformations are responsible for the two style groups of major folds. S_2, S_3 and S_4, however, maintain attitude through the zone and if separate deformations are involved then an extra, major phase has to be invoked. Such an approach, appears unnecessary when the slide could have brought together somewhat dissimilar structural domains. The orientation of the D_1 X direction (maximum elongation) appears to be similar in the two blocks (see Figures 7.27 and 7.28) and acts to link the zones structurally.

The major fold above Terminal Peak on the profile (Figure 4.35) is a multiple hinge type similar to Figure 7A of Ramsay (1967), with an axial surface dip of between 30° and 40° to the south. The limbs are essentially planar and an extensive planar zone is developed in the closure. Within this closure the two intermediate-sized, rounded isoclines appear to be large scale symmetric (M) folds and disharmony must be considerable here. Large variations in bedding thickness occur in this region. The outer portion of the double hinge fold is in massive, thick-bedded (0.5 to 1 m) quartzite which outcrops as large regular slabs. The inner portion is more thinly bedded (0.1 - 0.4 m) with many phyllite selvages between the quartzite layers. This multilayer type produces zones with much minor folding.

Below the thrust and to the north of Frankland Peak, a major close D_1 fold occurs in quite thickly-bedded quartzite. The immediate area of the closure is a zone of minor folding, which, with the lack of a 'marker' horizon, makes any measurement of layer thickening impossible. Around this fold, S_1 cleavage is fanned from about 310.31 on the overturned, steep limb to about 268.69 on the right way up more gently inclined limb. As the axial surface dips at about 45°, the fanning is of the order of 15° on either side. Pronounced fanning of S_1 is a feature on all scales in this portion of the Franklands and appears in all stereoplots (Figures 4.38 sheet 1, sheet 2; 4.40 and 4.43). In the Frankland Peak area above the slide, S_1 cleavage fans are found in the closures of the rounded major isoclines but S_1 parallels S_0 on the limbs. The fanning of S_1 must in part be due to later deformations though the extent of S_1 fanning and refraction produced solely in D_1 is impossible to estimate.

Between Greycap and Terminal Peak, minor D_1 folds are ubiquitous and younging based on truncated cross-bedding is often readily determined. The asymmetry of the minor folds and facing are congruous with the major folds as mapped out on bedding surface attitudes. If two major deformation events were responsible for the large pre-D_4 structures of the region, then incongruous minor fold relations and complex reversals of younging might be expected. The overall picture is of major events in D_1 and D_4 with sporadic development of D_2 and D_3. Only one occurrence of a proved post D_4 structure is known apart from rare kinks. A cleavage (S_5?) with attitude 124.53 cuts S_4.

Minor D_1 folds vary considerably in style within quartzite units from this region from open to isoclinal and may be strongly flattened or virtually 1b types (Figure 4.44 a, b). S_1 is usually well developed in F_1 closures but may disappear on the limbs. Though S_1 is a common structural element it is locally overprinted and in the field obliterated. Some of the more massive quartzite does not display a fabric in outcrop. S_1 in thin section (46246, 46247, 46248, 46250, 46251, 46252, 46257, 46294) is a variably developed, dimensional preferred orientation of detrital quartz grains and fine recrystallised mica. Intragranular plasticity appears to be the main deformation mechanism during D_1 and there is only limited evidence for solution transfer in this event (44366). Recrystallisation is variable, ranging from zero to about 80%. The microscopic nature of S_1 is little different on either side of the slide and in several traverses the change from $S_0//S_1$ to $S_0 \wedge S_1$ was documented (44371 to 44370 and 44374). S_1 is particularly well developed in pebbly grit horizons (44257) and should provide worthwhile material for further strain studies.

In the Franklands region, east of the Saddle, some difficulty was found in recognising different cleavages in the field. All surfaces may take on a grain appearance, particularly those that occur as dissolution seams in very slightly micaceous quartzite. In some instances two low angle 'grain fabrics' were taken to be one surface until thin section studies were made (e.g. 46246). In other situations a strong dimensional preferred orientation of 5 mm sized particles was overprinted in surrounding finer layers by a weak dissolution event and the latter cleavage was the only one recognised in the field (44257). Even if a strong grain alignment is present it is often equalled or surpassed by an apparently weak dissolution event, in terms of field expression. Weathering along the zones of mica in the seams must etch the exposed surface to a greater degree than along dimensional preferred orientations. Other problems in cleavage recognition occur where S_1 has a steep dip and lies close to the S_2 or S_4 orientation. This is the position on the right way up limb of the major close fold to the north of Frankland Peak. Specimens from this limb have a strong dimensional preferred of quartz grains (S_1) at about 30° to S_0 (46247) and two sets of dissolution seams a low angles to this direction. These latter are interpreted as S_2 and S_4. On the overturned limb two cleavages develop, one dipping about 30° and the other about 70° to the south southwest. They appear to be spatially segregated but in detail the former is found to be S_1 and the latter a dissolution seam type of S_2 which obscures S_1 at times. The subdued expression of S_1 and the similarity of S_1 and S_2 in outcrop caused some problems in pinpointing the thrust in the initial mapping. S_2 had to be proved by overprinting on first folds or D_1 boudins and then extrapolated to other localities. S_2 overprinting a platyness parallel to S_0 had to be interpreted cautiously as the bedding parallel fabric may have been sedimentary.

4.3: Post D_5 Minor Structures

Kink-bands, which involve the rotation of 2 to 8 cm zones of earlier cleavages, are found in all sub-areas but are not abundant. Though approximately one hundred measurements of kink-plane attitudes were taken, the wide aerial extent over which data were collected has limited interpretation. In any one region, kink bands tended to fold S_4 which was little affected by later events. Readings on kinks within each sub-area could thus be compared without too much influence of initial variability of the kinked surface.

The Starfish region (sub-area 4) was sampled in some detail to characterise the kink-band geometry. Lithologies in this region are marked by well developed planar fabrics which favour the formation of kink-bands. A complex pattern was found (Figure 4.45), essentially demonstrating the presence of two generations of pairs of conjugate kinks. The general trends of these two pairs are:

010 - 190	Group 1	D/N
330 - 150	Group 1	U/N
050 - 230	Group 2	D/E
085 - 265	Group 2	U/E

Following E. Williams (1970), kink-folds are distinguished by quoting the displacement of the pre-existing cleavage, reading from left to right, whilst noting the general viewing direction. In one outcrop it was noted that a kink from group two folded a kink from group one, therefore, establishing overprinting relationships. Though most of the readings taken of kink-planes on The Starfish can be related to these two events, several do not fall into these groupings. It is indicated that three, or possibly four, kink producing deformations occurred.

For the rest of the Franklands/Wilmots area, only clear examples of the most frequently developed conjugate kink-pairs have been plotted (Figure 4.46). Group one on The Starfish can be directly correlated with a conjugate pair developed around the Gordon Dam by rotating one pair through 20°; a rotation which is required by the regional change in hinge line bearings (see Figure 4.47). The generally east/west trending kink-bands on Mt. Sprent and the Wilmots are probably equivalent to group two of The Starfish.

A prominent conjugate kink-pair in the Cleft Peak to The Lion area has two sets; U/N trending 025-205 and D/N trending 215-035. In this region the statistical fold axis has been rotated by 70° from the north/south trend at the Gordon Dam. A rotation of this amount would bring the kink-bands of Mt. Sprent and the Wilmots into approximate alignment with the kinks between The Lion and Cleft Peak. Between Frankland Saddle and Terminal Peak, an U/N set trends 360-180 and a conjugate D/N set trends 020-200. The regional fold axis has undergone a further 26° rotation from The Lion area and this again can account for the change of trend of the kink-plane attitudes. Some variation in kink plane attitudes would be expected if the two conjugate kinks events were superimposed on an already rotated sequence of metamorphosed rocks. A consideration of the stresses required to produce the kinks at the northern end of the mountain chain, indicates that this stress pattern would not produce the conjugate kink pair seen in The Lion area if the major rotation had already taken place.

The generally north/south conjugate pair at the Gordon Dam can be shown to have been rotated through about 20°. The generally east/west kinks in the north appear to have been rotated to northeast trends and then to north/south strikes in the south of the region. In the latter case rotations are clearly correlatable for the last 30° of movement though somewhat uncertain for the first 60°. Patterns for each sub-area are complex and the above data taken in combination with other published work suggests the possibility of four generations of small scale kinks. The rotation of the whole Frankland and Wilmot mountain chain, probably pre-dates the Middle Cambrian (see Section 4.5). The evidence presented above demonstrates that two conjugate kink-pairs pre-date this whole chain displacement. S.J. Williams (1976) found north/south and west southwest/east northeast kink sets in Precambrian rocks which he could correlate on orientation grounds with structures produced during the Middle Devonian Tabberabberan Orogeny. These latter structures may well be sporadically developed throughout the Franklands/Wilmots area, thus giving rise to locally complex kink-plane orientations.

4.4: Discontinuous Structures

Very few fault planes have been directly observed during the course of this study. In the Southern Wilmots several prominent valleys may follow fault lines but displacements must be small, unless they are of a trace slip nature. Major lithological contacts may be slightly offset but poor exposure and gradational boundaries combine to blur definition of such features. There is a marked tendency for these prominent valleys to run approximately at right angles to the crest length of the major ridge. It is possible that they are etched out along master joints which parallel the ac plane (Price, 1966, p. 114).

Exposures created by engineering workings in the Gordon Dam-Knob Quarry area allowed twenty four readings to be taken on fault surfaces; the majority of the faults had perpendicular separations of less than 3 metres. Three groupings of the readings were evident:

i) 160 - 340, 10 readings on mainly reverse faults, dipping between 50° and 60° to the northeast.

ii) 120 - 300, 10 readings on mainly normal faults, dipping on average 70° to the southwest.

iii) 015 - 195, 4 readings, faults dip west at about 80°.

A similar pattern is described by Andric, et al., (1976) who extensively characterised the fracture patterns in the works area during a study of the engineering geology. Low angle (less than 30° dip) faults with irregular fault planes were noticed in several outcrops but not measured. Invariably these surfaces were undulating with amplitudes of ½ to 1 metre over wave-lengths of 2 to 3 metres. Andric, et al., (op. cit.) described a low angle fault in the Gordon Dam foundations which required careful investigation for its possible effect on dam stability. To the western edge of the Knob Quarry runs the most noticeable fault of the region in terms of effect on outcrop pattern. This fault belongs to the 340-160 trending group. In the quarry area, fold enveloping-surfaces are nearly horizontal as is layering in the thick quartzite unit. The vertical separation on this large fault is between 15 and 40 metres. The uncertainty in the displacement occurs because the layering may in part be stepped down by asymmetric, D_1 folds.

Joints have not been sampled in this work but it was noted that ac joints are very common and exert a considerable influence on the topography of the main mountain ridge.

4.5: Synthesis of Major Structures

Several gross features of the Franklands/Wilmots area become apparent when stereoplots of bedding readings for each subdivision are plotted on a topographic map of the whole region (Figure 4.47). π axes show a gradual swing of trend from 264.20 at the southern end to 002.00 at the northern extremity. Between Terminal Peak (264.20) and Redtop to Cleft Peak (289.10) there is a departure from the progressive change where the Greycap area has a 310.20 π axis. This variation has fairly sharp boundaries on either side and is a kink like fold of all the structures in that area. The time of formation of this kink with respect to the major rotation is uncertain. All structures, up to S_5, studied follow the major swing of trends and it is, therefore, post S_5 in age. F_4 axial surface traces most clearly define this trend but F_1 traces also illstruate the swing Figure 4.48). No closely defined upper limit may be placed on the age of the major rotation within the thesis area. To the northeast similar strike changes have been demonstrated by the Department of Mines to be pre-upper Middle Cambrian (unpublished mapping results). Drag folding on a major transcurrent fault may have been responsible for rotating the major structures of the region (Corbett, 1970).

Along the mountain chain, π axes tend to plunge at low angles to the west, northwest or north with a slight reversal around the junction of the Wilmots and Franklands. The greater part of the Wilmots have no detectable plunge to the larger structures. Some doubt exists as to the exact angular relation between fold axes of the major deformations D_1 and D_4, and also D_2/D_3 where these become important. The π girdle patterns are basically simple (Figure 4.47) displaying fairly regular cylindrical distributions. This appears to be largely a result of essentially having one event as the dominant fold producing episode in each region. Other examples include major

D_1 isoclines with poorly represented closures being folded by major D_4 folds (e.g. Redtop to Cleft Peak); such overprinting would give simple π girdles even if the events were non-coaxial. Where D_3 produces major folds (Tombstone Hill, The Starfish, Mt. Sprent to Gordon Dam) the patterns are invariably complex with either two cylindrical girdles or a cylindrical/conical mixture. Low angle cross folding appears to be present in the Gordon/Serpentine area and may persist along the whole chain though it is more apparent towards the north where the opportunity exists for better definition. At Terminal Peak in the south, major, tight, D_1-folds are folded by major upright D_4 folds. Most of the sampling was of the first folds but their π axis is not well defined because of the preponderance of long straight limbs at low angles to one another. D_1 and D_4 may be slightly oblique but this cannot be proved satisfactorily.

D_1 and D_4 produced the most widespread effects on all scales while D_2 and D_3 only achieved prominence in the central and northern portions of the region. Within the mapped area, D_2 folds are no larger than second order in size but may expect to be found on a larger scale over a greater distance. Consistent asymmetry of the largest D_2 folds indicates sub-horizontal movement from the present day west. D_3 gives rise to a first order fold on The Starfish but elsewhere is on a smaller scale. Again the largest D_3 folds show consistent asymmetry involving in this event an S shape. Structural sequences can generally be analysed in terms of large scale rotation of D_1 - D_3 features during D_4 (Figure 4.49). Prior to D_4 enveloping surfaces were probably flat-lying to gently inclined as D_2 and D_3 would have produced local rotations of a sequence folded by tight to isoclinal recumbent D_1 folds. Pre-D_4, F_2 axial surfaces would have dipped moderately to the west and F_3 axial surfaces to the east. D_4 produced near upright folds with steep axial surface dips to the east and rotation of D_1 - D_3 structures. Minor later events overprinted D_4 and a major strike-slip fault probably caused rotation of all the above structures such that north/south trends became east/west in part.

The large size of the D_1 folds and the dominance of bedding parallel first generation planar fabrics, suggests that the earliest event involved the production of nearly flat-lying nappes. Prior to the rotation that affected all D_1 to D_5 structures, the first fold axes would have been approximately rectilinear with the Mt. Sprent/Gordon Dam region representing that trend which now parallels the north/south axis of Tasmania. On the south, southwest or west dipping limbs of major D_4 synforms the D_1 folds are all upwards facing whilst on the north, northeast or east dipping limbs they face downwards. The original pile of first folds which now parallel the north/south long axis of the present day Precambrian outcrop in Tasmania, would have faced towards what is now the east. The major transport direction during D_1 would therefore have been from the present day west. This is the same conclusion as reached by Powell (1969a) but his evidence came exclusively from folds here ascribed to D_2. D_2 of this work demonstrates repeated large scale overriding from west to east and D_1 - D_2 may be the products of a continuous process of crustal movements.

CHAPTER FIVE

MINERALOGY AND PRESSURE/TEMPERATURE CONDITIONS DURING DEFORMATION

5.1: Introduction

The whole region examined in this study is clearly zoned in terms of metamorphic grade but it is not possible to give good definition of lines marking the incoming of certain key phases such as garnet. A zone characterised by the presence of garnet is found on the east side of Detached Peak (Figure 5.1). It is an elongate, north/south aligned, belt of schistose pelite running up to the Strathgordon/Knob road. Immediately west, phyllite, especially at the Gordon Dam, is quite glossy but the grain size rapidly decreases towards Mt. Sprent. From this latter point southwards to Terminal Peak, all pelitic rocks are fine grained and appear slaty in places, particularly in the far south. The micaceous quartzite of The Starfish tends to be coarse grained but again grain size diminishes quickly to the south. Highest grades were attained in the eastern portion of The Bell/Detached Peak area (sub-area 3) whilst the bulk of the thesis area experienced lower grade conditions. Garnetiferous rocks also are found to the north northwest of the Gordon Dam on the Hamilton Range (Scott, in Spry and Banks, 1962, p. 117), and several kilometres to the south southwest of the Scotts Peak Dam, approximately 10 km southeast of Terminal Peak.

Mineralogical aspects as a whole are more consistent with greenschist facies conditions with albite being the only feldspar recorded. Chloritoid is recorded from a small number of localities. White micas are typically phengitic and epidote minerals are common in metabasites in which actinolitic amphiboles are abundant. Muscovite-chlorite pairs are well distributed whilst calcite occurs sporadically. All the above features are characteristic of the greenschist facies according to Turner (1968). However, following Turner's classification scheme, almandine is rare in the greenschist facies yet it is common in the higher grade zone of The Bell and Detached Peak. Biotite has not been recognised anywhere in the Franklands or Wilmots and chlorite is only found in very small quantities. The typical pelite is a qtz + phengite + tourmaline + zircon rock with minor traces of chlorite. From the metamorphic viewpoint most effort in this study was put into the strip of country between Mt. Sprent and the Serpentine Lookout (Figure 5.1) where conditions were thought to show the greatest variation.

5.2: Textural Analysis

Garnets vary from being sieve textured (40689) through poikiloblastic (40690) to euhedral forms with no inclusion (46241). Within the garnets small inclusions of quartz often display a strong dimensional preferred orientation. These internal trails (S_i) are for the most part straight suggesting a helicitic (Spry, 1969) preservation of an earlier surface. Where inclusion trails exist they invariably become curved towards the margin of garnets demonstrating that the last phase of garnet growth was syn-tectonic. Garnet growth was post-D_1 and, on the basis of cleavage relations in specimens 40689 and 40690, synchronous with the earliest part of D_2. Internal and external trails are nearly continuous or only show slight amounts of rotation with respect to one another. Garnet growth may well, therefore, have continued to the end of the second cleavage forming event. Chlorite porphyroblasts in 40690 show similar features to the garnet, with generally straight inclusion trails at an angle to the external S_2 but with curved marginal trails which pass virtually without break into S_2. Porphyroblastic albite has been noted in slides prepared by the Hydro-Electric Commission (V.E. Thompson, pers. comm.). Albite contained only straight inclusion trails and was presumed to have overgrown S_1. Small (0.05 mm), inclusion free, garnets in specimens 7211034 and 40666 are strongly wrapped by a near penetrative second cleavage and have 'pressure shadow' zones associated with them. Garnet growth, in these examples, ceased before D_2 or early in D_2.

Quartz in inclusion trails is smaller than the general external quartz indicating that immediately post-D_1 the pelitic rocks were still fine grained. In the pelites, a general stage of matrix coarsening must be correlated with the pre- and syn-D_2 garnet growth. The degree of coarsening and compositional readjustment of minerals varies from area to area. In lower grade areas these features seem only important in crenulation cleavage zones (see discussion of mineral chemistry variation between microlithons and S_2 cleavage planes - Chapter 3.5); at higher grades the processes are more pervasive. In the garnet zone pelite, phengite grains are found up to 1.5 mm long, perhaps averaging 0.5 mm. In differentiated schists, large phengite grains tend to show a very strong dimensional preferred orientation in mica rich layers parallel to S_2. Some stubby phengite up to 0.6 mm long is found in the mica domains at high angles to S_2, giving rise, locally, to a crude decussate fabric (46241). Obviously phengite growth continued after D_2, the coarse size of the S_2-parallel phengite possibly being in part a mimetic effect. In a garnet-schist (46305), phengite in differentiated mica rich layers (S_2) has in part grown at the expense of phengite in the quartz rich microlithons where it traces out microfolds.

Recrystallisation and growth of metamorphic minerals during or after D_3 is limited. Porphyroblasts of garnet, chlorite, albite or chloritoid do not show evidence for growth post-D_2. There is, however, the possibility that porphyroblasts may have continued to enlarge a little post-D_2 and overgrow the external structures. If S_i and S_e were continuous post-D_2, a small, post-tectonic, addition to the porphyroblasts may be undectable. In D_3- and D_4-microfolds of nearly pure mica layers, recrystallisation has led to the folds being segmented into small straight or slightly curved sections. Individual grains interlock and the pattern is indicative of the migration of high angle grain boundaries (Etheridge and Hobbs, 1974). The process is usually confined to high curvature regions at the closures of the micro-folds. Distinct new grains of mica are only rarely seen in D_3 or D_4 crenulated pelite, where they are confined to the crenulation cleavage planes.

Small phengite grains occur parallel to S_3 and S_4 in slightly micaceous quartzite. No direct evidence is available for the origin of these grains but no progressive sequence of rotation has been noted and a dissolution and reprecipitation mechanism may be responsible. Quartz is the mineral mostly affected by post-D_2 events and its recrystallisation in zones of high strain in D_3 and D_4 appears to be largely a strain induced feature.

Textural evidence points to a thermal climax at about the time of D_2 with the peak perhaps occurring slightly prior to D_2. Garnet, albite and chlorite porphyroblasts can all be shown to have developed prior to S_2. Temperatures were maintained at sufficiently high levels to allow continued growth of porphyroblasts during D_2; some individuals ceased growing during D_2, whilst some may have grown until a little after D_2. A tendency for some decussate overgrowth of differentiated D_2 crenulation cleavages shows a gradual waning of metamorphic temperatures. The subsequent thermal history is uncertain as temperatures were never again sufficiently high to produce porphyroblasts. Any peaks or troughs of the thermal regime cannot be detected. D_3 and D_4 may have occurred during a gradual and smooth decrease of temperature or either or both events may have been associated with an elevation of temperature after a relatively cool period. Whatever the thermal history, D_3 and D_4 certainly took place at lower temperatures than D_2.

5.3: Metamorphism of the Metasedimentary Rocks

The distinction between metasedimentary and meta-igneous rocks in the Strathgordon area is clear in the vast majority of cases. One exception to this statement is an amphibole-albite-chlorite-quartz schist at the Serpentine Lookout (Figure 5.1). As this rock is predominantly amphibole bearing it is described in the section on metabasites for convenience.

The following mineral assemblages have been found in the region:

1) quartz + phengite + tourmaline + zircon + opaques (haematite).
2) quartz + phengite + chlorite + tourmaline + zircon + opaques.
3) quartz + phengite + chlorite + garnet + tourmaline + zircon + opaques.
4) quartz + phengite + garnet + chlorite + albite + tour. + zir. + opaques.
5) quartz + phengite + chlorite + chloritoid + albite + garnet + zircon.

Analytical Procedure

The previously unpublished electron probe analyses that are presented in this chapter were carried out by Mr. B. Griffin at the School of Earth Sciences, the Australian National University. Details of the instrument and data analysis are given in an earlier study on the same area by Boulter and Raheim (1974).

Phengite. After quartz, phengite is the most abundant mineral. It shows a slight green colouration in thin section under plane polarised light and whole sections are pleochroic if a strong dimensional preferred orientation is present. Pleochroism is very marked in some thin sections. In the structural formula, the Si:Al ratio is greater than 3:1, hence the mica is classified as phengite (see Table 5.I). Mg and Fe^{II} have substituted for Al in the octahedral sites, thus accounting for the colouration. The electron probe analyses show that FeO varies between 2.5 and 5.5% and MgO between 1.5 and 3.5%.

Phengite is known to be markedly zoned on the grain scale (Boulter and Raheim, 1974) and this, therefore, requires detailed probe traverses to be made before effective use may be made of phengite compositions as P/T indicators. A very thin but pronounced marginal variation is attributed to a late event. Within the garnet zone, cores of phengite grains have very similar Si^4 values based on 11 oxygen (40667, 3.22, 3.29, 3.23; 7302043, 3.26, 3.23, 3.21). The latter specimen is chloritoid bearing and is likely to be much more iron rich than the former, yet despite this compositional difference both have similar Si^4 contents of phengite in the garnet zone. Near the Gordon Dam, a chlorite-calcite-quartz-phengite rock which must be very iron rich has phengite Si^4 values of 3.34, 3.35 and 3.42. According to Velde (1965, 1967), at approximately similar pressures, a range of Si^4 content from 3.35 to 3.24 reflects an increase of nearly 100°C in temperature. The work of Boulter and Raheim (1974) demonstrated the need for caution in the interpretation of such data and the necessity of combining textural analysis with chemical analysis. The Si^4 variations may be reflecting the regional metamorphic gradient from the Gordon Dam to the garnet locality on the road. However, in part, the values may reflect a sampling of D_1 mica in one area and

TABLE 5.I

PHENGITE ANALYSES

PHENGITE

	39166		46289 THREE SEP.GRAINS			46271 SAME GRAIN			40667		40667		40667				
	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	A(9)	A(10)	B(11)	B(12)	C(13)	C(14)	C(15)	C(16) Margin	C(17) Core
SiO_2	50.24	50.00	50.22	49.14	50.17	46.76	47.76	45.50	48.62	46.42	48.50	48.10	47.32	47.57	46.27	49.92	48.22
TiO_2	-	0.00	0.43	0.48	0.35	0.23	0.33	0.13	0.00	0.21	0.17	0.13	0.18	0.24	0.00	0.14	0.21
Al_2O_3	28.29	27.54	24.54	25.16	24.47	28.17	29.72	29.37	29.23	30.58	29.38	29.02	31.10	30.34	31.09	28.53	20.96
FeO	3.11	2.96	5.17	5.26	4.87	5.17	5.53	4.66	3.74	3.11	3.74	3.89	3.15	3.36	2.68	3.05	3.34
MgO	3.23	3.36	3.52	3.47	3.25	1.91	1.81	1.60	2.23	1.92	2.37	2.12	1.75	2.04	1.63	1.75	2.01
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	-	-	-	-	-	-	-	-
Na_2O	0.21	0.24	0.26	0.18	0.00	0.20	0.18	0.27	0.84	0.54	0.86	0.82	0.31	0.97	0.37	0.41	0.65
K_2O	10.88	10.65	11.17	10.96	11.45	10.25	10.93	9.98	9.68	9.31	9.72	9.57	10.05	9.49	9.24	9.06	9.84
Cr_2O_3	0.45	0.76	0.00	-	-	-	-	0.00	0.00	-	-	-	-	-	-	-	-
	96.41	95.51	95.31	94.60	94.55	92.69	96.26	91.53	94.35	92.29	94.76	93.65	93.85	94.01	91.29	92.86	95.84
	ATOMS																
Si	6.675	6.704	6.844	6.752	6.884	6.529	6.445	6.415	6.588	6.428	6.550	6.572	6.433	6.461	6.425	6.787	6.461
Ti	-	0.000	0.044	0.049	0.036	0.024	0.034	0.016	0.000	0.021	0.018	0.014	0.018	0.024	0.000	0.015	0.021
Al	4.429	4.352	3.942	4.064	3.958	4.635	4.726	4.880	4.668	4.970	4.676	4.673	4.983	4.856	5.088	4.572	4.889
Fe	0.346	0.332	0.589	0.604	0.558	0.604	0.624	0.549	0.424	0.358	0.423	0.444	0.358	0.382	0.311	0.347	0.374
Mg	0.641	0.671	0.716	0.711	0.664	0.398	0.365	0.336	0.451	0.395	0.478	0.432	0.354	0.413	0.338	0.354	0.401
Na	0.055	0.063	0.068	0.048	0.000	0.054	0.048	0.075	0.221	0.143	0.225	0.217	0.082	0.255	0.098	0.107	0.169
K	1.843	1.821	1.942	1.922	2.004	1.826	1.881	1.794	1.674	1.638	1.674	1.668	1.743	1.645	1.637	1.570	1.682
Cr	0.047	0.081	0.000	-	-	-	-	0.000	0.000	-	-	-	-	-	-	-	-
	14.036	14.022	14.145	14.151	14.103	14.070	14.122	14.064	14.026	13.995	14.044	14.021	13.970	14.036	13.898	13.752	13.998

Calculations of Structural formulae based on 22 oxygen. ANALYST B. GRIFFIN the sampling of metamorphic maxima mica in another. In a prograde sequence, with D_1 before the maxima, this would also give a trend of rising temperature. Boulter and Raheim gave 3.30 as the average Si^4 value for D_1 and 3.26 for D_2 (≡ the metamorphic maxima here). These values were best defined on the road section immediately east of the Gordon Dam, between that dam and the Serpentine. The lower average of Si^4 of 3.24 for the garnet zone D_2 micas appears to be reflecting the general regional zoning.

Velde's work (1965, 1967) has been criticised on the basis of the validity of the experiments, in that some runs were not reversed and starting products were not always appropriate. Also the application in some areas has been questioned because the apparent use of non-buffered assemblages (Guidotti, 1973). Raheim in several publications has demonstrated the validity of using Si^4 values in a wide range of whole rock compositions and metamorphic grades (Raheim and Green, 1974; Raheim, 1975, 1976). Within the present area phengite contains Mg, Fe, and Na in significant quantities. The terrain involves chlorite and albite bearing rocks which would buffer the Mg, Fe and Na component. Some specimens additionally contain garnet and chloritoid to further strengthen the case for a buffered assemblage. Given that Velde's experimental data are reasonably sound, the Si^4 content of phengite can be used in this region to monitor P,T conditions. At the present time there appears to be a growing concensus that potassic white-mica compositional variations are more sensitive to pressure fluctuations than thermal effects (Sassi and Scolari, 1974; Fettes, et al., 1976). However, in the Strathgordon area the Si^4 trends from lower grades into the garnet zone are totally compatible with Velde's (1965, 1967) predictions of response to increasing temperature. Both pressure and temperature, therefore, appear to influence the Si^4 content and though

Velde's curves (Velde, 1967) may not be accurately placed, relative positions are useful.

The value of 450±50°C given by Boulter and Raheim (1974) for the metamorphic climax in the Strathgordon area is in the general region of the transition between the greenschist and amphibolite facies. Such a transitional position is suggested by the regional mineralogy and further demonstrates the general validity of the P/T relations proposed by Velde (1967). The application of the b_0 method of Sassi and Scolari (1974) is extremely difficult in the Strathgordon area because of polyphase deformation. It is very rare to find mica grains relating to only one structural event. The typical situation involves three cleavages and markedly zoned mica. Any b_0 determination would only give an average of a markedly heterogeneous assemblage.

In order to use the Si^4 information from the Strathgordon rocks it is necessary to have information concerning the pressure during the peak of metamorphism. This is essentially taken from the work of Raheim (in prep) who has used, as a pressure estimate, the relationship between the Si^4 curves and the curves for the breakdown of Fe-chlorite and Mn-chlorite to give almandine and spessartine (Hsu, 1968). The garnets (specimen 40669) are almandine with 15-20 mol.% grossular, 4-8 mo.% spessartine and 3-5 mol.% pyrope components (Raheim, in prep). The iron rich composition suggests that the garnets were derived from Fe-chlorite. The relation of the breakdown curves indicate a 3-7 kb pressure bracket and Raheim (in prep) estimated P conditions of 4±1 kb for the metamorphic climax. This estimate was in part using mixed data of mica from the climax, but from a zone a little cooler than the garnet zone, together with the inferences about the garnet stability. By allowing for the effect of the non-almandine

TABLE 5.II

CHLORITE ANALYSES

CHLORITE

	39166		46271	40667					
				SAME GRAIN		SAME GRAIN		SAME GRAIN	
	A	B		A	A	B	B	C	C
SiO_2	26.33	26.89	24.52	24.69	24.88	24.73	24.78	24.73	24.44
TiO_2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al_2O_3	21.03	21.56	21.23	22.13	22.03	22.20	21.92	21.71	22.00
FeO	21.80	22.37	35.69	33.01	32.99	32.05	32.64	32.33	32.47
MgO	19.32	19.46	9.33	10.48	10.66	10.03	10.51	10.75	10.33
CaO	0.00	0.00	0.11	0.00	0.00	-	-	-	-
Na_2O	0.23	0.35	0.60	0.66	0.60	0.49	0.64	0.65	0.59
K_2O	0.00	0.00	0.09	0.00	0.00	0.00	0.08	0.00	0.00
MnO	0.15	0.18	-	0.00	0.00	-	-	-	-
Cr_2O_3	0.52	0.24	-	0.00	0.00	-	-	-	-
	89.37	91.05	91.56	90.97	91.15	89.50	90.57	90.18	89.83
				ATOMS					
Si	5.349	5.363	5.238	5.223	5.248	5.289	5.259	5.264	5.228
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Al	5.035	5.068	5.344	5.517	5.475	5.594	5.481	5.447	5.546
Fe	3.703	3.732	6.376	5.839	5.818	5.731	5.792	5.755	5.808
Mg	5.849	5.785	2.970	3.303	3.350	3.196	3.324	3.411	3.293
Ca	0.000	0.000	0.026	0.000	0.000	-	-	-	-
Na	0.090	0.135	0.248	0.272	0.246	0.205	0.265	0.269	0.245
K	0.000	0.000	0.024	0.000	0.000	0.000	0.021	0.000	0.000
Mn	0.025	0.030	-	0.000	0.000	-	-	-	-
Cr	0.083	0.038	-	0.000	0.000	-	-	-	-
	20.183	20.151	20.226	20.154	20.137	20.015	20.142	20.146	20.121

Calculations of Structural Formulae based on 28 oxygen. ANALYST B. GRIFFI components on Hsu's Fe chlorite → Fe-garnet breakdown curve and by taking Si^4 values from the same zone, a P/T estimate at the climax in the garnet zone would be 5±1 kb at 500±50°C.

Chlorite. The chlorite here is green and pleochroic with extremely low birefringence and is invariably a very minor component of pelitic rocks. Some specimens may contain 5-10% chlorite but in most phyllite it is in lower quantities than tourmaline. Raheim (in prep) has reported secondary chlorite but all the examples presented here are primary and probably produced before or during the metamorphic climax. Mg values (100Mg/Mg+Fe) of chlorite grains vary between 68 and 35 from rock to rock but within one specimen are uniform (Table 5.II). High values of MgO (19 wt.%) are only found in specimen 39166 which is likely to be a contaminated igneous rock. Chlorite from metasediments ranges from 7 to 11 wt.% MgO and have FeO contents from 32 to 36%.

Garnet. In the Strathgordon region, garnet is found as very skeletal forms (40689) and well shaped prophyroblasts free from inclusions (46241). It is a common mineral to the east side of Detached Peak and The Bell and may form up to 10 vol.% of any one specimen. The garnet is 70 to 80 mol.% almandine (Raheim, in prep) and the average meta-pelite, though perhaps low in iron, has the appropriate composition to produce this mineral.

Tourmaline occurs both as rounded and euhedral grains with the latter probably being derived from the former. Overgrowths on detrital grains can give a well formed appearance. Fine needle varieties are common in pelite and must represent total recrystallisation or perhaps crystallisation from dissolved tourmaline components. Analyses (Table 5.III) show the tourmalines to be iron-magnesian varieties between schorlite and dravite. Optically the grains are more related to schorlite being yellow-green in colour and strongly pleochroic.

TABLE 5.III

	ALBITE				TOURMALINE	
	46212		46289	46290*	41706	
SiO_2	68.81	69.52	70.73	67.60	35.65	36.04
TiO_2	-	-	0.00	0.16	1.14	.35
Al_2O_3	20.20	20.25	20.58	20.03	29.06	33.49
FeO	0.00	0.00	0.00	0.12	8.60	7.66
MnO	-	-	-	-	-	-
MgO	0.00	0.00	0.42	0.00	9.30	6.98
CaO	0.14	0.00	0.19	0.00	3.06	0.68
Na_2O	12.03	12.19	12.69	11.37	1.44	2.31
K_2O	0.18	0.14	0.25	0.12	-	0.00
	101.37	102.11	104.26	99.40	87.66	87.52

* Tas Uni. probe.

ANALYST: B. GRIFFIN

Chloritoid occurs as prismatic needles and is only found in the garnet zone. The maximum size recorded is 1.4mm x 0.08mm. The chloritoid is strongly pleochroic from blue to very pale blue with a faint tinge of burgundy in the interference colours. Grains are length fast. Chemical analyses (Table 5.IV) are similar to other published examples (Chinner, 1967).

Albite as porphyroblasts including S_1 has only been reported in one slide. A more common form is as widely dispersed grains which are of a similar size to the surrounding quartz grains. In the coarse (possible para) amphibolite of the Serpentine Lookout (46212), albite averages 0.5mm. Analyses (Table 5.III) show the albite to be nearly the pure end member with usually less than 0.2 wt.% of K_2O or CaO.

5.4: Metabasites and Related Rocks

Amphibolite is common in the surface and underground workings of the Hydro Electric Commission at Strathgordon and is found in several natural exposures in the northernmost portion of the area studied. South of Mt. Sprent and Detached Peak they are sparse with one example occurring south of Remote Peak on the Companion Range and another south of Greycap. Mapping by the Hydro-Electric Commission geologists has shown that the majority of amphibolite bodies cut bedding and several examples show well preserved igneous textures and mineralogies. In 46243, large partially altered pyroxene grains are surrounded by a plexus of well shaped laths which, however, are totally altered. One clear case of a cross cutting body (46290) gave a major-element analysis similar to that of tholeiitic dolerite (Table 5.VI). This example was similar to several analyses carried out by Raheim (1976) on amphibolite and eclogite occurrences 70 km to the north which were considered to be

TABLE 5.IV

CHLORITOID

	GRAIN 1			GRAIN 2	
SiO_2	24.65	24.62	24.24	24.69	24.77
TiO_2	-	-	-	-	0.00
Al_2O_3	41.71	41.57	41.02	41.29	41.66
FeO	28.25	27.44	26.44	28.75	27.05
MnO	0.00	0.14	0.00	0.13	-
MgO	2.69	2.28	2.51	1.67	2.44
CaO	-	-	-	-	0.00
Na_2O	0.51	0.57	0.43	0.58	0.76
K_2O	-	-	-	-	0.00
	97.21	96.73	94.64	97.10	96.62

SPECIMEN 46271 ANALYST B. GRIFFIN

meta-igneous rocks. One problematical amphibolite occurs at the Serpentine Lookout (46211 to 46213). It is concordant with the compositional layering in the host metasediment and, like those clearly of igneous derivation, has a high content of opaques. However, the MgO wt.% in the chemical analysis (Table 5.VI) is very high (~ 17 wt.%). Pre-metamorphic igneous rocks are widespread, but minor, constituents of the Tasmanian metamorphosed Precambrian. Spry has shown quite a wide range in their compositions but they are generally consistent with a derivation from an olivine-basalt magma type (Spry, Chapter 8 of Spry and Banks, 1962). All of Spry's analyses from discordant amphibolite bodies are much lower in MgO and the Serpentine Lookout amphibolite seems most likely to have been an impure dolomite initially. Specimens 39166 and 46300 came from one cross cutting body. Calcite and chlorite each compose about 40% of this rock with minor phengite, quartz and opaques accounting for the remainder. The unusual composition of this material (Table 5.VI) is best accounted for in terms of contamination of igneous material by resorption of metasediment.

The time of emplacement of the dolerite dykes in the tectono-metamorphic cycle is uncertain. The majority appear to have been intruded prior or during the metamorphic climax. Preservation of igneous textures may suggest a post- or late syn-tectonic position but invariably metamorphic amphibole has been produced. Such textures are best developed where amphibolite occurs in pelite, which would take up most of the stress difference, or in regions where D_2 is of low intensity. Metabasite emplacement can be either pre-D_1, syn-D_1, post-D_1 before the metamorphic climax, or even during the peak of metamorphism. They usually carry one good fabric and one minor crenulation whereas most pelite has one pervasive cleavage and at least two crenulations. On the road to the base of the Gordon

Dam, several transecting amphibolite bodies were exposed in the road cuttings. The larger body contained a weak cleavage in the S_3 orientation and a weak crenulation in the direction of the local S_4. A penetrative fabric was found in a smaller body parallel to the S_2 surface in the enclosing quartzite. A post-D_1, pre-D_2 age of intrusion is indicated here. Unfortunately, clear examples of this nature are rare. Though frequent in underground workings, the amphibolite bodies act as channels for water movement and become deeply weathered. Contact reactions due to intrusion and (?) metamorphism blur relationships and only scant evidence is available to place the meta-dolerite in the structural sequence.

The following mineral assemblages have been recorded in amphibolite:

1) amphibole + albite + chlorite + opaques.
2) amphibole + albite + epidote + chlorite + opaques.
3) amphibole + albite + zoisite + chlorite + opaques.
4) pyroxene + amphibole + albite + (Ca fels?) + epidote + opaques.
5) calcite + chlorite + qtz + phengite + opaques.

1) to 4) mostly have quartz as a minor accessory. 1) represents the probable metasediment from the Serpentine Lookout as well as some meta-igneous bodies. 5) is from the Gordon Power Station and is thought to be a contaminant of igneous and metasedimentary material.

Textures range from being completely igneous in nature to thoroughly recrystallised nematoblastic fabrics. The latter texture is found in transgressive bodies and the probable metasediment of the Serpentine Lookout which also contains a near penetrative crenulation where S_1 is almost obliterated. The more complex textural history of this rock type is indicative of its presence throughout the structural cycle. Some amphibolite occurrences contain two petrographic varieties of amphibole. 46291 is a totally recrystallised rock in which 90% of the amphibole is extremely pale green yet has a maximum extinction angle Z^c of 25°. The remaining 10% is strongly pleochroic from deep blue green to pale green and is often the kernel to a grain dominated by the pale variety. These features are very common in metabasites of the Scottish Dalradian from the chlorite to the garnet zone (Graham, 1974). 46218 has relict igneous minerals and a crude metamorphic fabric. One grain in 46218 is subdivided into three parts, with narrow transitions, from pyroxene to strongly pleochroic amphibole (deep blue green to yellow green) to faintly pleochroic amphibole (pale blue green to colourless). Generally the extremely pale amphibole is most widespread (46214, 46280, 46290, 46298, 46299) whilst moderately coloured blue green amphibole (Z^c, 29°) is found in 46212. 46290 and 46212 both have pure albite in which CaO and K_2O are both less than 0.2% (Table 5.III).

Electron probe analyses show amphibole to be calciferous (Table 5.V). This mineral group is complex in chemistry, and its major features are essentially dependent on bulk rock chemistry. Certain variations can, however, be related to metamorphic grade. The analysed amphibole samples are classified as actinolite and actinolitic hornblende according to the system of Leake (1968) using the parameters Si, Ca+Na+K, Mg/Fe+Mn+Mg from the structural formulae. 23 oxygen atoms have been used for the calculation of the structural formulae whilst assuming that all the iron is in the ferrous state. Undoubtedly the ferric state is represented in calciferous amphibole but probably not in large quantities. The whole rock analyses show this to be the case. Because iron is determined as FeO by the probe, all values quoted for structural formulae are slightly high (Stout, 1972).

TABLE 5.V

AMPHIBOLE ANALYSES

	46280 Actinolitic Hornblende			Actinolite 46290			Actinolitic Hornblende 46212				
	Grain 1	Grain 2	Grain 2								
SiO_2	51.85	50.88	52.14	54.29	54.18	55.15	52.04	51.96	49.43	52.76	51.59
TiO_2	0.10	0.00	0.00	0.00	0.00	0.00	-	-	-	-	-
Al_2O_3	6.10	9.42	6.48	3.55	2.58	1.68	5.80	5.68	6.64	5.11	5.85
FeO	13.20	15.04	13.28	14.31	15.33	13.52	9.97	9.75	10.35	10.14	11.26
MnO	0.13	0.16	0.13	0.14	0.00	0.12	-	-	-	-	-
MgO	14.29	12.55	14.34	14.94	14.59	15.78	16.56	16.74	15.36	16.61	16.03
CaO	10.31	9.77	10.50	12.01	11.59	11.73	10.47	10.84	10.17	10.03	9.93
Na_2O	1.52	2.06	1.63	0.83	0.85	0.46	2.20	2.00	2.37	2.57	2.56
K_2O	0.14	0.16	0.12	0.13	0.10	0.00	0.24	0.29	0.39	0.19	0.31
Cr_2O_3	0.11	0.00	0.00	0.00	0.44	0.30	0.32	0.42	0.56	0.41	0.44
	97.76	100.04	98.62	100.19	99.70	98.76	97.60	97.67	95.27	97.83	97.97
					ATOMS						
Si	7.46	7.20	7.44	7.66	7.72	7.85	7.42	7.41	7.28	7.51	7.38
Al^{IV}	0.54	0.8	0.56	0.34	0.28	0.15	0.58	0.59	0.72	0.49	0.62
Al^{VI}	0.49	0.77	0.53	0.25	0.15	0.13	0.40	0.36	0.43	0.37	0.37
Fe	1.59	1.78	1.58	1.69	1.83	1.61	1.19	1.16	1.27	1.21	1.35
Mn	0.02	0.02	0.02	0.02	-	0.02	-	-	-	-	-
Mg	3.06	2.64	3.05	3.14	3.10	3.35	3.52	3.56	3.37	3.52	3.42
Ca	1.59	1.48	1.60	1.82	1.77	1.79	1.60	1.66	1.60	1.53	1.52
Na	0.63	0.31	0.45	0.23	0.23	0.12	0.61	0.55	0.67	0.70	0.71
K	0.02	0.02	0.02	0.02	0.02	0.00	0.04	0.05	0.07	0.03	0.05
Cr	0.02	-	-	0.00	0.05	0.03	0.04	0.05	0.07	0.05	0.05
	15.42	15.02	15.25	15.17	15.15	15.05	15.39	15.39	15.48	15.41	15.47

Structural formulae calculations based on 23 oxygen   ANALYST B. GRIFFIN

Because of limited sampling, the present study can only indicate trends but appears to be a fruitful line to pursue in terms of relating the Strathgordon area to other well known zones and also for showing trends of rising metamorphic grade within the region. A full analysis should indicate the position of Strathgordon in relation to the Barrovian, Glaucophanitic and other facies series. Wenk et al., (1974) have noted the need for many analyses so that the means of various components can be compared within subdivisions of a region and with other terrains in the world. Significant standard deviations restrict the use of small samples such as presented here.

Amphibole of a given Mg content has been shown experimentally and in several metamorphic regions to become more tschermakite rich with increasing metamorphic grade (e.g. Graham, 1974). Amphibole samples from 46290, 46212, 46280 show such a trend with the latter being the highest grade matching its closest position to the zone of garnet growth in metasediment of The Bell/Detached Peak area. The data from the above amphibolite occurrences plot in the same field as analyses of amphiboles from the southwest Scottish Highlands, both on $Al^{\underline{VI}}/Si$ and $Na+K/Al^{\underline{VI}}$ diagrams (Graham, 1974). Though the range of values is much more restricted in the Strathgordon example, the similarities, particularly the alkali contents, suggest a similar pressure regime to the southwest Dalradian. Graham (op. cit.) has compared this latter area with the 'Barrovian' Haast Schists and the 'glaucophanitic' type of Shikoku and found it to be intermediate between the two. Load pressure during the metamorphic peak at Strathgordon is, therefore, indicated to be greater than in the intermediate pressure 'Barrovian' facies series.

Amphibolite samples 46244 and 46218 are from the far south of the area. They are both extremely fine grained with zoisite and chlorite predominating. Their mineralogy and relict igneous textures suggest a lower grade of metamorphism for this region which is consistent with the lack of any porphyroblasts in the adjacent pelitic rocks.

5.5: Discussion of Pressure Temperature Conditions

The majority of the following discussion will deal with the pre- and syn-D_2 metamorphic peak. The subsequent thermal history will be touched upon briefly. Chinner (1967) has recorded chloritoid-muscovite-chlorite-albite-quartz-garnet assemblages in the almandine zone of the Scottish Highlands but always in association with garnet-biotite schist. The metasedimentary assemblage 5) of this work was regarded by Chinner (op. cit.) as high greenschist or in terms of Turner (1968) low amphibolite facies. Chloritoid is found throughout the greenschist in rocks of appropriate composition and Turner (op. cit.) advocates the use of chloritoid as an indicator of low grade conditions. At higher grades, staurolite is considered to take over in iron rich aluminous pelite, though the simple notion of chloritoid/staurolite transition was questioned by Hoschek (1969) who pointed out that staurolite bearing rocks are closer to normal pelitic compositions than chloritoid rocks. Staurolite free assemblages must reflect temperatures less than 500°C (Hoschek, op. cit.) and 550°C is the upper stability limit of chloritoid. In the Strathgordon region, chloritoid is only found in the garnet zone and appears to be absent from the lower grades. This is presumably a feature of the P/T conditions; it is unlikely that appropriate compositions for the production of chloritoid only occurred in the higher grade regions.

An important feature of the region is the absence of biotite (see also Williams, S.J., 1976) whilst garnet is common in the schistose (coarse grained) rocks. A similar order of appearance of index minerals is noted in the Sanbagawa metamorphic terrain of Japan (Kurata and Banno, 1974). In Japan the propensity for garnet formation is attributed to higher pressure conditions in the glaucophanitic Sanbagawa belt than those of the Scottish Barrovian sequence. Mather (1970) has shown that the first appearance of biotite is a function of rock composition, particularly the Al_2O_3:Fe,MgO ratio. Specimens with low Al_2O_3 and high Fe,MgO are the first to develop biotite. With high aluminium bearing compositions the appearance of biotite is delayed until higher grades. Many Strathgordon pelitic rocks are low in iron (Table 5.VI) but this is not always the case. 40683 (Table 5.VI) contains nearly 7% iron and 46271 has large amounts of chlorite and chloritoid suggesting significant quantities of iron and magnesian. The pelite is not particularly aluminous when compared with world averages. Fe/Fe+Mg whole rock ratios fall in the range given by Kurata and Banno (1974) for the Sanbagawa and Dalradian metamorphics. Biotite always appears first in arkosic metasediment which is an unknown rock type in the Strathgordon area, but clearly the area of garnet growth is beyond the biotite zone. Korikovskiy (1973) shows that the formation of almandine garnet in the place of Fe rich chlorite can be regarded as the upper limit of the biotite zone of metamorphism. This is well exceeded at Strathgordon with no sign of biotite in the rocks. A more extensive investigation of compositional control is warranted but it seems likely that a sufficiently wide range of compositions is present to prefer an explanation in terms of P/T conditions being unfavourable to biotite growth (or more favouring garnet).

The metamorphosed Precambrian of Tasmania does show evidence of being a medium to high pressure terrain generally. Raheim's metamorphic

TABLE 5.VI

Chemical analyses of meta sedimentary and meta-igneous rocks - Analyst B.J. Griffin, FeO determinations by C.Boulter & B.J.Griffin

	46213	46211	46307	46306	46304	46305	46290	46303	46302	46301	39166	46300	40683	40709	46308
SiO_2	46.08	46.23	76.27	64.49	93.86	78.96	51.64	60.25	72.82	66.56	44.46	48.27	61.26	69.92	87.12
TiO_2	1.06	1.10	0.45	1.12	0.13	0.46	1.30	1.50	0.87	1.10	1.12	1.27	0.64	1.06	0.59
Al_2O_3	10.52	11.10	12.05	21.04	2.76	11.02	12.64	21.58	15.20	19.22	12.46	13.37	18.77	15.65	5.65
Fe_2O_3	1.87	2.27	0.74	0.38	0.09	0.88	1.75	0.16	0.00	0.00	1.59	4.45	2.83	0.30	1.89
FeO	9.60	8.89	1.71	0.58	0.58	1.69	9.10	1.85	1.60	1.14	8.66	8.79	3.74	2.22	0.00
MnO	0.17	0.17	0.08	0.00	0.00	0.03	0.15	0.00	0.00	0.00	0.14	0.22	0.27	0.00	0.07
MgO	17.14	17.20	1.85	1.97	0.93	1.03	8.71	2.53	1.33	1.72	8.83	8.90	1.36	2.01	0.85
CaO	6.86	6.42	0.01	0.08	0.00	0.04	8.95	0.00	0.00	0.01	9.75	7.87	0.05	0.02	0.00
Na_2O	1.28	1.04	0.25	0.00	0.01	0.41	1.65	0.01	0.01	0.12	0.04	1.34	0.08	0.02	0.00
K_2O	0.73	0.94	3.92	7.04	1.07	3.28	1.38	8.61	5.63	6.98	1.02	0.45	6.72	5.97	1.88
P_2O_5	0.08	0.10	0.04	0.06	0.01	0.01	0.11	0.02	0.01	0.02	0.09	0.10	0.03	0.02	0.06
Loss	4.61	4.55	2.63	3.34	0.56	2.21	2.62	3.49	2.53	3.14	12.23	4.97	4.30	2.82	1.93
	100.00	100.01	100.00	100.10	100.00	100.02	100.00	100.00	100.00	100.01	100.39	100.00	100.05	100.01	100.03

All analyses normalised to 100%-ignition loss

46306 40683, 40709, 46301, 46302, 46303 } Phyllite specimens - Gordon Dam area.

46304 Micaceous quartzite - Serpentine Quarry

46305 } quartzose schists The Bell
46307 } quartzose schists The Starfish

46300 39166 contaminated igneous rock

46290 Meta-dolerite

46211 amphibolite - Serpentine Lookout
46213 (Metasediment?)

46308 weathered quartzose phyllite

study some 70 km to the north (Raheim, 1976) indicated very high pressure conditions for another part of the same terrain. Kyanite has been reported in several localities (Turner, 1971; Raheim and Green, 1974). Further supporting evidence for a general high pressure terrain with high P/T ratios comes from the absence of granitoid activity of a syn- or post-metamorphic type. The metamorphosed Precambrian occurs in a belt approximately 270 km x 70 km in extent with areas of later cover. Throughout this area granitoid activity is low and no proved Precambrian examples have been recorded. Such a feature may be a function of erosion level but over the whole terrain once deeply buried metamorphic rocks are fairly common. Zwart (1967) has pointed out that granitoids are rare in high pressure terrains becoming more frequent towards low pressure situations. Possibly the high P/low T conditions in Tasmania were not able to produce conditions of partial melting of the lower crust. If the orogen were a continental margin type, then thinner crustal conditions may also have combined with fairly high P/low T gradients to further reduce the likelihood of producing granitoid magmas.

The baric characteristics of the Strathgordon area are further confirmed by the presence of pure albite in (46289) and very close to (46212) the zone of garnet growth. This again indicates a higher load pressure than the classical Barrovian series (Crawford, 1966) as was supported by the limited amphibole data.

In the highest grade zone temperatures are likely to have been less than 500°C on the basis of chloritoid and staurolite behaviour. Using semiquantitative information about the relations between facies on a P/T plot, a temperature of between 400 and 500°C is suggested. The metamorphic peak a few kilometres to the west of the garnet zone was given at 450°C by Boulter and Raheim (1974) using a pressure estimate of 4 kb. Somewhat higher pressure values may be valid but these would only increase the temperature marginally according to Velde's (1967) phengite curves. Noting the absence of staurolite, reasonable P/T conditions for the metamorphic peak are 450±50°C at 6 kb±1 kb in the high grade zone.

D_3 of Boulter and Raheim (1974) is equivalent to D_4 of this study. Analyses of micas texturally related to this event, suggest a marked drop in temperature by this time. Little is known of the conditions during the present D_3 except to note that mica underwent limited recrystallisation and no new garnet or albite grew post-D_2. D_4 of the present study has been equated by Raheim and Compston (in prep.) with a partial resetting of Rb/Sr systems approximately 150-200 m.y. after the peak of metamorphism at 800 m.y. before the present. The drop to the D_4 temperature may not be a progressive cooling curve and a trough may have developed in the curve before again rising in response to a later thermal/structural event. The D_1-D_5 sequence is, therefore, considered by Raheim and Compston to be two separate tectonic-metamorphic cycles (Tobisch et al., 1970).

CHAPTER SIX

THE DEFORMATION OF SEDIMENTARY STRUCTURES

6.1: General Considerations

In any discussion of a metasedimentary terrain which involves the study of sedimentary environments, the evidence for primary structures and their state of preservation should be presented. These considerations have more weight in this instance, as the modification of sedimentary structures is used as important information for the analysis of large-scale tectonic structures. Several authors (Whitten, 1966; Hobbs, Means and Williams, 1976) have emphasised the importance of presenting proof of a primary origin because "sedimentary structures are commonly modified, obliterated or simulated by deformation and recrystallisation" (Christie in a review of 'The British Caledonides', Oliver and Boyd, 1963). In the present work, it is also shown that some purely sedimentary structures could be mistaken for tectonic structures. A further complication arises from the possibility of a sequence being regionally deformed before lithification is well advanced (Davies and Cave, 1976). The latter situation might be expected to produce features normally attributable to both sedimentary and tectonic processes and the interplay of such features could be complex.

The clastic dyke occurrences (Chapter 2) provide a wide range of geometrical relations between bedding and various cleavages. Because of the rising prominence of dyke/cleavage relations in support of the tectonic dewatering theory (Maxwell, 1962), some considerable time was devoted to their study. From the outset, it appeared that the deformation of sedimentary structures was poorly understood mainly through a lack of attention. Hence, the remarkable variety of dyke/cleavage relations were used to the full. The following sections attempt to demonstrate that without combining studies on sedimentary structures and cleavages, both of which have been thoroughly researched, little idea can be gained of their expected geometrical configurations after strain. If such work is not carried out, erroneous conclusions can be obtained concerning cleavage/soft sedimentary structure relations.

6.2: The Tectonic Dewatering Hypothesis

A dominantly mechanical origin for slaty cleavage was proposed by Maxwell in 1962, the primary agent being the tectonic dewatering of piles of rapidly buried sediment. Several interpretations of Maxwell's original paper are possible but it appears that the important concept involved the production of abnormally high pore pressure during the deformation of unconsolidated material. High pore pressures which did not exceed lithostatic pressure would reduce friction between grains thus allowing deformation to rotate the grains fairly readily (see Tullis, 1976). Maxwell was, however, considering the situation where pore-fluid pressure exceeded lithostatic for the formation of the clastic dykes which he stated were parallel to cleavage. It is this latter notion which led to the modification proposed by Powell (1969b) who envisaged that when the lithostatic pressure was exceeded, water, carrying pelitic material, streamed out of the sediment along fine channel ways. Mica was, by this mechanism, rotated to become aligned in the channels (the cleavage planes) and also deposited in the channels. The tectonic dewatering mechanisms give the basis of slaty cleavage, and the preferred orientation so derived, may be enhanced by later strain (Powell, 1972b). Powell (1969b, 1972a, 1972b, 1973) has been responsible for amplification and refinement of the theory and has given further details of structures thought to be characteristic of this mode of cleavage formation.

Maxwell's hypothesis (1962) suggested a radically different environment for the formation of slaty cleavage to that previously considered. No longer was cleavage formation thought of as a metamorphic feature, but it was stated to depend upon the unconsolidated nature of thick piles of sediment. Since Maxwell's paper, there have been several contributions supporting tectonic dewatering on a regional scale (Moench, 1966, 1970; Braddock 1970; Clark 1970; Lynas 1970; and Bishop 1972). Supporters of tectonic dewatering contend that soft-sedimentary structures and slaty type cleavage are related. The recognition and theoretical basis of tectonic dewatering rested heavily on the parallelism of clastic dykes and first cleavages, and such relations were believed to indicate the unconsolidated nature of many sequences when deformation occurred (Maxwell 1962, p. 287; Powell 1972a, p. 96 and 108; Alterman 1973, p. 34). In some cases the evidence for tectonic dewatering may be quite flimsy (LaFountain, 1975) and may even be based on a solitary outcrop of clastic dykes sub-parallel to a slaty cleavage (Bates, 1975).

Prior to 1974 few studies had been directed toward determining the directional convergence of surfaces of sedimentary structures and later cleavages during deformation. Jones (1937, p. 271-272), Bielenstein and Charlesworth (1965) and Nickelsen (1973) have described clastic dykes parallel to cleavage and considered them to be sedimentary in origin. Elliston (1963, figure 49 and p. 229) has shown some nearly planar clastic dykes in cleaved rocks, but his description of relations to tectonic structures is unclear. Also, Smith and Rast (1958) discussed penecontemporaneous clastic dykes in the Dalradian succession but made no reference to secondary structures.

6.3: Clastic Dykes and Tectonic Structures in the Franklands/ Wilmots Area

The discussion of dyke/cleavage geometrical relations will be concentrated on the first two clastic dyke localities described in Chapter 2.2: ii. The evidence for a pre-cleavage, soft-sedimentary origin is initially proposed on the basis of (1) the origin of several sets of dykes from beds that were subsequently eroded. (2) the presence of many examples of intersecting dykes.

6.3: i) Tectonic Structures at Locality One. Throughout the road section (Gordon River Road, Figure 2.8), the compositional layers dip steeply to the east (Figures 6.1 and 2.9). The part of the outcrop dealt with in detail is structurally simple, but 250 m to the east, there are strongly flattened, coupled isoclines (D_2 folds with an axial plane crenulation cleavage). These folds (Figure 6.2) occur in an alternation of chloritic schist with thin (0.05 to 0.2 m) quartzite layers, and they show vergence changes over a distance of 10 m, which indicate an F_2 synform. No sedimentary structures are preserved in these rocks. In this eastern exposure, a transecting crenulation cleavage (S_3 of Figure 6.2) dips about 50° to 80° to the east and is associated with the weak development of a new layering by metamorphic differentiation. Though this cleavage is S_3 on the scale of the road section, it is probably equivalent to S_4 on the regional scale.

The transecting cleavage can be traced almost continuously back to the clastic dyke occurrence where it may be confused with S_2 which here is found at a low angle to $S_0//S_1$. S_2 has a slaty appearance in outcrop (40658, 40660) and even in thin section, its crenulation nature can be difficult to determine (40658, 40661, 40666), though in some areas it appears to be absent. S_3 is generally a well-spaced (1 cm) crenulation in pelitic rocks but is very patchily developed. S_1, parallel to S_0, is usually preserved as a continuous structure (Dennis 1972) in the silty layers and is similar to that illustrated by Powell (1969b, Figure 3) and Williams (1972b, Figure 7D). Second folds in this immediate area have rarely been recognised but appear to be microscopic, attenuated, isoclinal folds with axial surfaces nearly parallel to the compositional layering (40690).

Within the area where the clastic dykes are located, the units face east, though a little farther east there is an indication of overturning. No mesoscopic folds are seen in this portion of the outcrop, but first folds on a larger scale could be responsible for the change of younging. The asymmetric scour and fill structures (Figure 2.10 a) were probably deformed in D_1 and are equivalent of S vergence folds when viewed towards the south. Garnets are found in this section but the specimens generally have a superficially undeformed or very low grade appearance.

6.3: ii) <u>Relations of Clastic Dykes to Cleavages at Locality One</u>.
Figure 2.13 illustrates the form of the clastic dykes as seen in a section perpendicular to the layering in an area where S_1 is parallel to the layering and the effects of S_3 and S_2 are negligible. Features like these in tectonically undisturbed rocks have morphologically comparable folds generally attributed to differential compaction between the dyke and the host mud. In Figure 2.13 the dykes are relatively undistorted where they pass completely through silty layers as there they would undergo a similar degree of compaction to that of the matrix.

The contorted form of these dykes must in part be due to D_1 because the direction of maximum shortening during this event would be parallel to that which operated during the compaction of the sediment. Hence D_1 would have accentuated the compaction folds and thus account for the extreme folding and disruption of some intrusions (40640, 40641). Some of the high angle dykes display in detail an accentuated form of ptygmatic folding which is often expressed solely as an irregular margin (ringed area, Figure 2.13) and reflects excessive shortening along their length (Figure 6.3).

The folded form of clastic dykes in undeformed sediment is usually considered to be entirely the result of compaction (Truswell, 1972, p. 583) though in this case measurements on the dykes will give the sum of compaction and tectonic shortening. Measurements along the middle line of dykes from Figure 2.13 usually give a range of percentage shortenings from 40% to 70% with the thick dyke in the top left hand corner giving a value of 20%. Even in parallel folds estimates of deformation by this method will be too small because of layer-parallel shortening (Hudleston, 1973, p.212). In addition, orthogonal thickness measurements (Ramsay, 1967) around several folds indicates an average of 55% flattening. Extreme total strains are indicated but unfortunately a precise solution for the total (or even minimum) shortening involved is made impossible by the small size, complicated form and variable composition of the dykes. This last point is probably the most difficult to overcome as the dykes not only vary in composition from one to the other but also along the length of an individual intrusion. The thickest dyke referred to (Figure 2.13) is very rich in pelitic matrix and hence would tend to deform in a much more homogeneous manner than some of the thin, coarser grained, quartz-rich dykes.

In marked contrast to the dyke forms, the sill-like bodies are fairly planar, though again in detail some examples of irregular margins might indicate strongly compressed compaction folds (Figure 6.4). The more regular form must be a result of their initial orientation close to the XY plane of both the compaction and tectonic strain (D_1) ellipsoids.

In some cases the transecting cleavage is very strongly developed in generally pelitic horizons which contain silty clastic dykes (40652, 40655, 40656). The form of these dykes is seen in Figures 2.11 and 6.5 where they are very close to the S_2 direction. In Figure 6.5 the angle between the dyke and S_2 is about 10 - 15°.

The exact three-dimensional form of the dykes is often difficult to determine though in several examples the intersection of a dyke with a compositional layer has been plotted (Figure 2.14). These intersection patterns are often simple as the hinges of both the compaction and tectonic folds are parallel to the layering. Irregularities are often cuased by uneven surfaces of the specimens. Two basic patterns are found, the first being more or less a hexagonal form (Figure 2.14 a, b, d), the second showing a strong preferred orientation (Figure 2.14 c, d, e). The linear intersections are usually within fifteen degrees of the S_1/S_3 intersection lineation indicating that the approximately planar forms of Figure 2.11 are sub-parallel to S_3. Because they have all the characteristics of the clastic dykes mentioned above, the crudely hexagonal patterned types are considered to be conjugate dyke sets, in the manner of Cook and Johnson (1970), rather than mudcracks.

6.3: iii) <u>Tectonic Structures and Relations to Clastic Dykes at Locality Two</u>. The phyllite in the Cleft Peak to Greycap region shows complex variations in the intensity of development of the different tectonic surfaces. Without overprinting evidence it is sometimes difficult to prove to which generation a particular surface belongs. Locally zones of a penetrative cleavage are noted which in thin section shows no trace of a crenulation nature.

This surface is taken to be S_1. S_4 in this area varies from zero development to quite an intense crenulation which at times is near penetrative but always maintains a near vertical attitude. In other zones of intense S_4 development, in association with clastic dykes (46297), earlier cleavages are preserved in thin section. The local preservation of S_1 at an angle to bedding at this locality might be expected because Cleft Peak is a second-order closure of a first fold.

In outcrop the clastic dykes appear to be perfectly parallel to S_1 (Figures 2.15, 2.16) but in large thin sections and slabs the situation is seen to be more complex. Figures 6.6 and 6.7 were taken from the outcrop illustrated in the sedimentology chapter (Figure 2.15) where the dykes were planar; they clearly demonstrate that the apparently planar form in outcrop is a simplified view. In Figure 6.7 most of the dykes are parallel to the cleavage, but occasionally they form tight to isoclinal coupled-folds with the cleavage axial planar. The situation illustrated by Figure 6.6 is more involved, with approximately planar dykes lying in the cleavage but having extremely irregular margins with frequent indentations (see area circles). Clastic intrusives at a high angle to the cleavage, and also at an angle to the compositional layering, are severely contorted; the axial planes of the folds being parallel to the cleavage.

Figure 6.8 shows sections of these features in the profile section of the fold (part a) and also parallel to the bedding (part b). The predominantly planar but slightly irregular form of the dyke structures contrasts markedly with the interconnecting, strongly folded branches. The quasi-planar forms lie almost in the cleavage as determined by their intersections on the two orthogonal sections (Figure 6.8). In Figure 2.16, the cleavage is refracted at the contact of the quartzite with the phyllite and the planar dykes faithfully follow this change of direction. This is similar to the example of dyke refraction figured by Powell (1969b, Plate 1, Figure 3).

6.3: iv) Discussion of time of formation of the clastic dykes and the present form of the dykes. In the cases described (particularly localities 1 and 2), a pre-cleavage liquefaction origin for the clastic dykes and sills is preferred because the first cleavage involved is usually, though in many cases only slightly, cross-cutting. All examples studied preserve examples of dykes, in the plane of bedding, which are at a high angle to cleavage despite, at times, a very strong outcrop appearance of dykes parallel to cleavage. The evidence shows that sedimentary structures, indicative of an unconsolidated state, were all entirely formed before the first major deformation event (D_1) and that these structures were variously strained depending upon local conditions (for example, initial shape, orientation, size, rheological properties of the rocks involved and contrasts between layers, etc.). The complex patterns of Figure 6.6 are thought to be essentially similar to those of Figure 2.13 from the Gordon Road and others in undeformed terrains (e.g. South Africa, Truswell, 1972; Von Backstrom, 1975). Though Powell (1969, p. 2592) has discounted rotation of sedimentary clastic dykes during strain, mainly because of the high degree of deformation that was considered necessary by theoretical considerations, it is proposed that there has been a strong convergence of orientations of sedimentary and tectonic structures during deformation in situations where bedding and cleavage are at an angle. Convergence is used to express the movement of linear and planar sedimentary and tectonic features towards the XY plane of strain such that they will come to lie nearer a common orientation as deformation proceeds

(Flinn, 1962, p. 424). Exceptions to this are that in progressive pure shear, structures parallel and perpendicular to the maximum shortening direction, will undergo no rotation but will be respectively shortened or elongated (Ramsay, 1967). Convergence can be simply demonstrated graphically (see below).

Localities 1 and 3 show a certain alignment of clastic dykes in a second or later cleavage direction whereas the present geometry of the Cleft Peak examples is regarded as being largely a result of strong deformation of sedimentary structures during D_1. Some other localities show marked alignment of clastic dykes in S_4 (46297). Quartzite layers in the Cleft Peak first folds often have profile geometries of quite strongly flattened parallel folds. Also the folded inter-connecting branches of intrusions from Figures 6.6 and 6.8 and specimen 40601, which lie at a high angle to S_1, give shortenings, perpendicular to the first cleavage, based on arc length measurements of 40%. The orthogonal thickness method shows that these folds where regular enough to measure have flattening values of 45 - 55%. Hence homogeneous tectonic shortening of the pelitic units, perpendicular to S_1, could reasonably be expected to well exceed 50%. Two examples of clastic dykes in undisturbed sedimentary rocks were taken from published figures (Figure 6.9) and graphically subjected to a 60% homogeneous plane strain through 20% increments. The final state is very close to that observed in Figure 6.7. This however, will only be an approximation to the mode of deformation of the clastic intrusions. A certain amount of body rotation of layers probably occurred as demonstrated by the folded form of the interconnecting structures of Figures 6.6 and 6.8. Also nothing is known at all of changes of dimension parallel to the hinge lines of the folds. Using the above construction as a guide it is felt that a combination of homogeneous flattening and body rotation could account for the irregular dyke margins, internal structures and near parallelism of dykes and first cleavages seen at Cleft Peak. A similar process may also explain the ptygma development in high angle dykes and irregular structures of the low angle offshoots in the Gordon Road examples (Figures 2.13 and 6.4) where tectonic and compaction shortening directions are parallel. Truswell (1972, p. 581) has shown that dykes at 25° or more to the bedding are invariably folded by compaction. Considering the large strain indicated for Figure 2.13 these low-angle, folded offshoots could become considerably modified during the tectonic strain producing the irregular margins noted. Such features could also be readily brought into alignment with S_1.

No estimate can be made of the quantity of shortening involved in the production of the crenulation cleavage S_2 seen in Figure 2.11 from the Gordon Road but it is considered that it will be less than the minimum inferred for D_1 at Cleft Peak. Even though, in sections perpendicular to the bedding, the dykes are sub-parallel to S_2 (Figure 2.11) and linear patterns are often close to the S_1/S_3 lineation orientation, it is thought that originally linearly arranged sedimentary structures and the strike of S_2 and S_3 were approximately coincident. The clastic dykes in Figure 6.5 probably showed accentuated compaction folds after D_1 and possibly were sub-perpendicular to the compositional layering and hence have been modified to a considerable degree. Also the strain associated with S_2 must have been sufficient to unfold compaction folds in some specimens (Figure 6.5) where the dykes are planar and sub-parallel to S_2.

40645 (Figure 2.14 d, e) shows two contrasting dyke/compositional layer intersection patterns on either side of the specimen. The strongly linear case is very close to the S_1/S_3 intersection lineation.

Its association with the branching pattern suggests coincidence in strike of the linear pattern and the third tectonic surface rather than marked convergence of orientations of sedimentary and tectonic structures during deformation, though the post-D_1/D_2 form of the intersections is uncertain. Some alignment of dyke intersections with the stretching directions of D_1 and D_2 probably occurred, but the near parallelism of the linear dyke intersections and S_1/S_3 lineation is fortuitous. The main reorientation during D_2 would be the change from a statistically perpendicular relationship between dykes and bedding (with S_1 parallel) to a situation where the dyke/S_0 angle is similar to the S_2/S_0 angle (c.f. Figure 2.11 with Figure 2.13).

Cook and Johnson (1970, Plate 3a) show a strong alignment of clastic dyke intersections with a weak inclination direction and give evidence to suggest that this is a result of a weak tectonic stress acting during sedimentation. The majority of their linear intersections were parallel to ripple marks but a relation to current action could not be supported. Donovan and Foster (1972, Figure 3) have demonstrated, in unstrained rocks, that oriented linear subaqueous shrinkage cracks are parallel to adjacent ripple crests and discuss the possibility of a relation to the palaeoslope of the basin of deposition or to an incipient directed tectonic stress. Situations such as these may reflect the fundamental tectonic control on basin shape (and hence current distribution), incipient stress directions and later more penetrative structures thus accounting for simple geometrical relations between tectonic and sedimentary structures (e.g. Furness, et. al., 1967, p. 145).

6.3: v) <u>The Tectonic Dewatering Hypothesis in the light of this and other Recent Studies.</u> The most significant evidence for the tectonic dewatering of slate is the presence of intrusive clastic dykes considered to be parallel to the first cleavage (Powell, 1972a, p. 108), which is taken to imply contemporaneous formation. Powell stated that liquefaction of sandy beds will take place early in the dewatering, and that pelite-bearing water may continue to stream out of the sediment, thus accounting for slight divergences of dykes and cleavage. Several examples of clastic dykes, in areas purported to show dewatering, have marked angles between cleavage and some dykes, and dykes are also occasionally strongly folded (e.g. Powell, 1969, Figure 2 and Plate 1, Figure 1). In Maxwell's (1962) type area, several authors have made accurate measurements of dyke/cleavage dihedral angles. J.B. Epstein (pers. comm.) has measured the outcrop of Figure 4A of Maxwell (1962) where a tabular dyke was shown as being parallel to the cleavage. In this example the dyke is 8° from the dip direction of cleavage and 20° from the strike of the cleavage. Geiser (1975) has shown that cleavage-dyke pairs have a mean dihedral angle of 14° with one outcrop giving 86°. It is considered, on the basis of the present study, that convergence of sedimentary and tectonic structures gives the best explanation of these relations. A further study of the Martinsburg type area (Beutner, et. al, 1977) has demonstrated on the basis of deformed calcite veins, that slaty cleavage in this region formed in lithified pelite and is unrelated to dewatering.

The mineral orientation mechanism in the type tectonic dewatering region has been studied from the view-point of mica fabric both by optical and X-ray methods (Holeywell and Tullis, 1975). The conclusion of this work is that mechanical rotation whether by tectonic dewatering or at high temperatures, cannot have been responsible for the preferred orientations seen in the slates. The fabric is considered to have formed by solution and recrystallisation processes.

From the examples quoted, it appears that, in slate belts, the angle between the cleavage and the dykes is quite low in sections at right angles to bedding. Slate belts commonly involve upright, gently-plunging folds with vertical stretching directions and many exposures might be expected to be along a c joints which approximate XZ sections. Hence the commonest exposures would equate with the highest strains and the lowest angular relations between features such as sedimentary dykes and cleavage. In all situations, bedding will be a lower strain plane and in areas of open folds will approximate the YZ plane. A complete study of dykes in slaty terrains should involve the measurement of the attitude of the dyke (Borradaile and Johnson, 1973 and Borradaile, 1974a) not just is apparent relation to other structures in one plane (see Geiser, 1975). An interesting example of varying dyke/ cleavage angles with different attitudes of exposures is given by Nickelsen (1973). In sections at high angles to both bedding and slaty cleavage, clastic dykes appear to be parallel to the cleavage. Initially this geometry was assumed to illustrate a soft-state cleavage genesis but examination of the dyke-bedding intersections showed flattened polygonal patterns with angular differences between dykes and cleavage. A nearly identical situation occurs in the slate belt near the Scotts Peak Dam, Southwest Tasmania (404203, Figure 1.2). In sections at high angles to bedding and cleavage the dykes are commonly sub-parallel to cleavage (Figure 6.10). When the bedding surfaces are examined, polygonal intersections appear (Figure 6.11a) and suggest a sedimentary dessication origin. The polygonal forms are, however, found in areas where bedding has a uniform attitude and where dykes intrude in both structurally up and down directions. Forceful injection of liquified sandstone is demonstrated, possibly triggered by earthquakes. A similar pattern of polygonal dewatering channels has been described by Johnson (1977) from little deformed sediment in North Norway. In the Scotts Peak district strain varies considerably and many examples of only slightly modified polygons are found along with varying degrees of preservation of diagenetic compaction folds (Figure 6.11 b). The flattening associated with the well cleaved outcrops, must have produced planar dykes from originally contorted specimens. Such a wide range of strain states is particularly useful in proving a sedimentary origin for the highly deformed examples and should be sought in all studies of dyke/celavage relations (Williams, D.M., 1976).

Fyson (1975) has figured clastic dykes sub-parallel to the second tectonic surface in biotite schist. Such a cleavage could not have a dewatering origin and amply demonstrates the alignment of dykes in cleavage during deformation.

The dewatering theory has been used to explain many features of cleaved rocks for which other mechanisms have been offered. Compositional layering parallel to first cleavages in lithicwacke and similar rocks has been interpreted as the result of intrusion of pelite-bearing water during cleavage formation. Williams (1972b) has given evidence that metamorphic differentiation is the cause in examples examined by him and has emphasised the production of compositional domains parallel to slaty cleavages in low-grade rocks. There is now a large number of investigations which support Williams' contention and many authors have dominantly ascribed tectonic layering to the pressure dissolution mechanism (Nickelsen, 1972; Durney, 1972; Geiser, 1974; Groshong, 1975; Durney, 1976; and Rutter 1976). Powell (1972a) regarded the fanning of cleavages as being best explained by dewatering. Cleavage with this mode of formation is proposed to develop as a planar feature virtually at one point in time during the deformation, and it is therefore, modified subsequently by continuing rotation. Cleavage fans will, however, form by inhomogeneous deformation of fold closures no matter how the original planar structure was produced.

Oertel (1970) showed that the observed strain in a birds-eye tuff from the English Lake District is sufficient to account for mechanical rotation of platy minerals to give a strong dimensional preferred orientation. Tullis and Wood (1975) and Tullis (1976) have shown a similar relationship for slate from Wales. The work of Holeywell and Tullis (1975), however, places constraints on these proposals because in the Martinsburg Formation they discounted mineral rotation as a mechanism. The conflict may be resolved by slate forming in many different ways.

The present work does not deny the possibility of dewatering cleavages, and local cases have been well documented by Williams et. al. (1969) in slump sheets of tectonically undeformed, thin-cover sequence. Similar features have been recorded by Tyler (1972, p. 549) and Moore and Geigle (1974); Corbett (1973) showed that cleavages formed in such a mobile fashion tended to be wavy in form. However, evidence for the operation and feasibility of the dewatering mechanism throughout basins several kilometres thick rather than on the scale of individual beds will have to be considered critically.

The geometrical relations between clastic dykes and cleavage assumes great importance when viewed in terms of its bearing on the evidence for tectonic dewatering. The above discussion has demonstrated that, with commonly observed strains, convergence of tectonic and sedimentary structures does occur to such an extent that dyke/cleavage relations cannot be used to support the dewatering hypothesis. Careful examination, in all reported examples, has shown that dykes are not parallel to cleavage and there is no substantial evidence to support Maxwell's 1962 proposal.

Alterman (1976) has recently considered that the tectonic dewatering mechanism would be expected to give rise to variations in dyke/cleavage angles. If this is the case there would be no unequivocal evidence in the rocks themselves to demonstrate the existence of a dewatering stage during the production of a regional slaty cleavage. Powell (1976) has argued that in some tectonic environments (trenches), incipient cleavage arises by tectonic dewatering and by extension he thus suggests that many slates could be initiated this way even if there is no direct evidence. Geiser (1976) considers the dewatering hypothesis "impossible to verify". In a detailed assessment of Maxwell's type area of the Martinsburg Slate, Groshong (1976) has demonstrated that deformation of sedimentary clastic dykes best explains their present geometry.

On the basis of dykes transected by cleavage (even if slightly) and features such as mud-crack origin (Nickelsen, 1973) or generation from near-surface beds that were eroded, a pre-cleavage origin can be proved for many dyke examples. Their close geometrical association with cleavage is a result of strain accompanying the development of the cleavage bringing about convergence. It is, therefore, concluded that the necessary parallelism of dykes and cleavage to prove the dewatering hypothesis has never been demonstrated satisfactorily.

6.4: Primary Deformation of Cross-Bedding and its Simulation of Tectonic Folds

In very well exposed areas it is possible to study the intersections of foreset beds on the upper erosional surfaces. These lineations are invariably curved and at some angle to the tectonic, rectilinear lineations (commonly a cleavage - bedding intersection). In a similar fashion, ripples may be proved to be sedimentary in origin, rather than mullions, by the bifurcating form and/or angular relation to tectonic lineations. The majority of cases are, however, not so clear and several complicating factors cloud the distinction between sedimentary and tectonic features. Initially it was feared that disrupted first-generation fold closures could give rise to apparently cross-stratified units which in particular would resemble herring-bone, cross-bedding. The tectonic situation would show opposite senses of younging if fully preserved and this has not been observed in the Franklands. Before the present investigation, proved examples of sedimentary structures were quite rare in the Tasmanian 'metamorphosed Precambrian' and thus caution was exercised against the ready acceptance of cross-lamination as being pre-tectonic.

Primary deformation (prior to lithification) can result in oversteepening of strata; the resultant geometry may appear to be part of a tectonic fold with one limb faulted out or attenuated. Tobish (1965) has recognised primary deformed cross-bedding in high grade tectonites and such features can clearly be recognised in the Franklands where the first tectonic surface (S_1) is at an angle to bedding. The primary nature of the oversteepening is particularly obvious when flattening associated with the cleavage formation, has slightly reduced the high angle cross-bedding dihedral angle. Unfortunately S_1 is most often parallel to bedding and thus many of the pre-lithification folds in cross-beds, described by Allen and Banks (1972), Hendry and Staufer (1975), Pettijohn and Potter (1964, Plate 110) and Jones (1962), may be mistaken as being solely of tectonic origin. Hendry and Stauffer (1977) have recently considered this problem but did not discuss the implications for structural analysis in detail. Near isoclinal primary folding of cross-bedding, where complete overturning of the foresets has occurred (Allen and Banks, 1972, Figure 2B), is morphologically identical to many tectonic folds. The layers are also thickened at the closure. If a cleavage is superimposed parallel to the overall layering, then it will be axial planar to the deformed cross-bedding further strengthening the similarity to tectonic folds. Even with high flattening associated with the cleavage, the deformed cross-bedding will not change its basic style. With cleavage parallel to bedding, examples with partial overturning of foresets (Allen and Banks, 1972, Figure 2a, c, d) may be mistaken for tectonic folds with thrust out limbs. The major problem appears to be the extension of the number of first generation tectonic structures by the inclusion of examples of penecontemporaneous deformation of cross-bedding. As most set thicknesses are less than 20 cm, a size cut off may be used where all larger intrafolial isoclines are most likely to be tectonic. The stable shelf environment further supports the size criterion as slump folds involving many beds are unlikely to form in this situation.

Several dangers arise in structural analysis if soft sedimentary folds are taken as tectonic. If such features occur throughout a cross-bedded unit which is affected by second or third order, asymmetric, parasitic D_1 folds, then the first fold may be assigned, incorrectly, to a later generation particularly if S_1 is poorly developed or obscured by later crenulations. This problem arises immediately upstream of the tailrace outlet for the Gordon Power Station. At this point, isoclinally folded cross-bedding occurs in the 30 metre long common-limb of a D_1 coupled fold. In the common limb, several examples of oversteepened cross-bedding can be traced into overturned foreset bedding, still preserving the upper erosion surface. In this example, S_1 is ill-defined and the presence of isoclines in the common limb may have led to the larger asymmetric fold being ascribed to D_2 or a later deformation. A more problematical situation occurs below the outlet structure at the Serpentine Dam, where small scale isoclines are found in cross-bedded quartzite. Some features suggest folded, truncated, foreset-bedding, but a strong fabric parallels the layering and the origin of the folds is equivocal.

6.5: Tectonic Deformation of Cross-bedding and Mechanisms of Formation of Large-scale D_1 Folds

This section will investigate the well exposed first folds of the Eastern Frankland Range which are found in virtually pure quartzite multilayers and where considerable contrasts in fold style occur. Cross-bedding and strain studies have been carried out on the major close fold (dihedral angle 32°) found to the north of the Frankland Saddle-Frankland Peak ridge (see Figures 4.35 and 4.36). The results of this analysis will be compared with the major isoclines that lie structurally above the close fold and use will be made of pertinent evidence from other localities.

Cross-bedding sampling stations were located on either limb of the close fold and at another site in a comparable structural position to the overturned steeply dipping limb. The unwinding procedure was that of Norman (1960) assuming that the fold plunge was due to a rotation about an axis at right angles to the fold hinge line. As the fold plunges are low (20° to 10°) it seems unlikely that a different rotation axis will affect the results markedly (Cummins, 1964). Norman's method (op. cit.) assumes a flexural slip folding mechanism (Ramsay, 1967, p. 392; Donath, 1962) and is in error for folds which have been significantly modified by homogeneous strain. At the outset it appeared that the major fold under discussion would approximate the flexural model as sedimentary grain fabrics are not markedly altered and cleavages are only moderately and non-uniformly developed over most of the fold. At sample site one, on the gently inclined right way up limb strain is low (see Chapter 7) with a moderate cleavage and 50 m up the sequence an intense platy fabric (S_1) is associated with only about 35% shortening across the cleavage.

As internal deformation is generally low the restored cross bedding orientations must be close to those of the undeformed state. The most common original current direction is from the northwest with other modes at northeast, southeast and southwest (Figure 2.7). Taking four representative cross bedding planes from each of the modes on Figure 2.7 their pitches on a north/south vertical plane were determined (represented in sketch form in Figure 6.12 b). This four layer sequence was folded schematically according to the flexural flow model where simple shear is distributed through the layers (Figure 6.12 b). On the gentle limb (sample one, 41182-70973) the dihedral angles between the foresets and topsets should decrease for the northwest/northeast currents and increase for the southeast/southwest currents (Ramsay, 1967, p. 492-498). The reverse is predicted for the steep overturned limb. Mean values for the dihedral angles of various modes are given in Table 6.1 where the behaviour of the northwest/northeast modes from one limb to the next follows the flexural pattern well. At sample point one the dihedral angle for the northwest set is decreased and at sample two (41188-71017 to 41182-70983) it is increased relative to the original value which, therefore, must have been about 28°. Average dihedral angles in a similar rock type have been quoted by Swett et. al. (1971) at 25° but with two modes at 25-30° and 10-15°. By taking large samples it is considered that a reasonable average is being attained. Unfortunately direct comparisons from one limb to the other of southwest/southeast mode dihedral angles is limited by the samll sample at site one. However at site two and three (41688-71039) the southwest/southeast sets are all below the estimated original angle of 28° which is the situation required by the flexural model. Because layering is well developed in the quartzite under discussion, deformation during folding must have been taken up by slip along the layering as well as slip distributed through the beds.

Ramsay (1967, p. 501) has produced curves to relate the changes in dihedral angles of obliquely inclined planes to the dip of the folded layer, the angle between the fold hinge and original strike of the inclined planes (α_0) (see Ramsay, 1967, p. 492, Figure 9-1), and the initial dihedral angle (δ_0). Assuming δ_0 to be about 28°, the other variables can be determined from Figure 2.7 and Table 6.1. Comparisons are given in Table 6.1 for predicted values of dihedral angles for a flexural model and a 50% shortened flexural fold with plane strain (Ramsay, 1967, Figure 9.16C and 9.19). The actually observed values of mean dihedral angles for this close fold do not fit either case but, considering the behaviour of the curves at high limb dips, the most likely test fit would be for an intermediate situation with approximately 25% homogeneous bulk strain.

The long limbed, fairly angular nature of the fold would indicate that tangential longitudinal strain, though possibly an important factor at the closure, would be in a very minor role on the limbs where the samples were taken (Ramsay, 1967, p. 400). Further strain studies in the closure would be required to determine the effect of tangential longitudinal strain but it does appear that, on a gross scale, flexural folding with a minor superimposed homogeneous strain can account for the fold structure. In the field the hinge zone is patchily exposed and an examination of the change of layer thickness in this region proved impossible. Also the profile section could not be reliably used for this purpose as no easily recognised layer was present for measurement. The three dimensional strain analysis (Chapter 7) indicates that layer parallel shortening (Huddleston 1973) was not significant in the formation of this particular fold.

Strain on a smaller scale is variable as suggested by the marked and rapid changes in cleavage intensity, and by the presence of rare parasitic folds. Several of the parasitic folds were well exposed with common cross bedding and one, a few tens of metres to the south of the main closure was chosen for further analysis (Figure 6.13). The northwest/northeast modes were common and on the gently inclined limb gave a mean dihedral angle of 19.3°. Two values from the same sets on the steep limb gave significantly different dihedral angles of 72 and 68°. With limb dips of 40°, $\alpha_0 \simeq 30°$ and $\delta_0 = 28°$ the measured values follow closely the curves of a 50% flattened flexural fold (Ramsay, 1967, Figure 9.19). As the α_0 values are determined from a flexural slip unwinding process they are obviously incorrect in this moderate strain situation; from a consideration of likely shortening directions it would seem that the α_0 values were higher than 30° originally. Nevertheless, the high dihedral angles on the short common limb of this parasitic fold demonstrate a locally increased strain, though quantification is inexact.

TABLE 6.1

SAMPLE	CURRENT MODE	MEAN DIHEDRAL ANGLE	NO. OF SAMPLES IN GROUP	FLEXURAL MODEL, DIHEDRAL ANGLE	50% FLATTENED FOLD MODEL, DIHEDRAL ANGLE	α_0
ONE	NW	23.8°	27	20°	5° - 10°	0° to 60°
TWO	NW	32.8°	9	47°	35°	45°
THREE	NW/NE	32.4°	17	65°	50°	0° to 20°
TWO	NE	35.2°	10	47°	35°	45°
TWO	SW	22.0°	12	20°	8°	45°
ONE	SE	17.2°	5	50°	38°	45°
TWO	SE	13.1°	8	20°	8°	45°
THREE	SE	21.5°	8	18°	15°	0° to 20°

Table showing mean values of dihedral angles to various modes at the cross-bedding sampling sites.

α_0 is the approximate angle between the strike of the restored foresets and the hinge line.

Sample locations: ONE, 41182-70973

TWO, 41188-71017 to 41182-70983

THREE, 41688-71039.

CHAPTER SEVEN

THE MEASUREMENT OF STRAIN

7.1: Outline of the Strain Analysis Project and Objectives

Considerable attention has recently been given to the methods of strain analysis involving populations of objects with a range of initial axial ratios and variable long axis orientations. Sedimentary fabrics ranging from those of oolitic limestone to coarse 'grit' have been studied by several techniques and, hence, the material of the present study, the quartz arenite of the Frankland Range, was within the variation that had been previously analysed.

In the material of this area, initial markers of dust inclusion trails around original detrital grains were occasionally clearly preserved and the compositional homogeneity of quartz grains and silica binding agent indicated conditions whereby the strain of the markers should closely approximate the true tectonic strain of each whole specimen. In grits several clast compositions predominate and each group with its own ductility has to be considered separately and the heterogeneities so introduced complicate the assessment of the tectonic strain for the whole rock (Dunnet, 1969). Ooliths often have relatively rigid cores which in reported examples influence the measured strain to such an extent that bulk tectonic flattening gives constrictional values for the strain from the oolith measurements (Tan, 1974).

Quantitative measurements on the amount of change of shape in Frankland Range quartz arenite assisted the investigations listed below.

1. The palaeocurrent analysis which had been carried out in folded and cleaved rocks, for it was important to know the order of magnitude of tectonic strain at the sampling sites in order to allow for this in the current reconstruction.
2. Fold mechanisms.
3. The field expression of cleavage in this rock type related to amounts of strain.
4. The relationship between quartz c axis fabric patterns and the amount of strain.

At the outset of the strain analysis program the most obvious problem was the almost universal presence of two or more deformation events at the hand specimen or thin section scale. In order to minimise these complications it was decided to restrict the study to the most massive quartzite zone i.e. the Frankland Range east of Frankland Saddle. This region had the most commonly preserved sedimentary features and here superimposed deformations were less intense than further north. See Table 7.1 for notation used in this Chapter.

7.2: Published Methods for the Analysis of Strain with Complex Initial Fabrics

Pre-deformation, non-spherical objects have been used as strain indicators for many years with the first comprehensive study being that of Cloos in 1947. Cloos measured oolith axial ratios in the strained state and took their mean as being the tectonic strain ratio. This in effect ignores the non spherical nature of undeformed ooliths and obviously would involve error if the initial ellipticity were marked. Such difficulties were recognised by Cloos who noted that the particle long axis orientations became much less scattered as deformation increased i.e.

TABLE 7.1

Terms used in relation to strain analysis methods

Rf	axial ratio of ellipse in deformed state
Ri	axial ratio of ellipse in undeformed state
ϕ	angle between ellipse long axis and maximum principal extension direction in deformed state.
θ	angle between ellipse long axis and maximum principal extension direction in undeformed state or reference direction (often bedding trace).
R_s	tectonic strain ratio
X,Y,Z	axes of tectonic strain ellipsoid X>Y>Z
R_{XY} etc	tectonic strain ratio in XY plane.

fluctuation decreased. He was however unable to incorporate this parameter in his analysis but came close (Cloos, 1947, p. 861) to developing the method published by Elliott in 1971. Cloos stated (1947, p. 861), " . . . if the diameters of nonspherical undeformed ooids were plotted in a co-ordinate system, a sphere would result due to the random orientation of the eccentricity".

Most of the latest advances have involved attempts at separation of tectonic strains from pre-tectonic shape factors, such as, depositional fabrics and compaction influences. Many studies now aim to give values for tectonic strains as distinct from total strains and also to measure compactional strains. This latter aspect is certainly in its infancy and the assumptions involved often appear invalid. The welding deformation in the Llwyd Mawr Ignimbrite, in Wales, was considered by Roberts and Siddans (1971, p. 292) to have been superimposed, presumably homogeneously, on a random shape fabric. Flow during the emplacement of this unit, however, may have given rise to a fabric prior to welding. Compaction strain in the majority of examples must be very heterogeneous with a high tendency for the shapes of markers to be insensitive e.g. quartz grains and pebbles in conglomerates. Oncoids and ooliths may undergo some limited change of shape whilst accretionary lapilli might be expected to show the greatest response to compaction. Certainly the situation is much more complex than the strain patterns considered by Sanderson (1976) though his work is invaluable as it clearly indicates the type of information that should be given in reports of strain analysis projects.

All the recently developed methods of strain analysis involve the measurement, in a plane, of the deformed particles' axial ratios (Rf) and their long axis orientations with respect to some reference direction (ϕ). A basic assumption is that strain has been homogeneous and a planar cleavage at the scale of investigation is usually taken to indicate that this holds. The method introduced by Ramsay (1967) which uses cleavage as the reference direction was modified by Dunnet (1969) to allow for more practical application to distributions that were initially close to random. Dunnet and Siddans (1971) advanced this technique to analyse non-random sedimentary fabrics essentially by unstraining along pure shear strain paths, in steps, until the distribution was most symmetric with the closest relationship between the bedding trace and the vector mean of the ϕ values being achieved. This will give the strain ratio if the initial fabric was symmetric about the bedding trace. For rocks with no equivalent of a bedding trace Roberts and Siddans (1971) modified the unstraining procedure such that the position of maximum symmetry determined the strain value.

A second method of strain analysis presented by Matthews, Bond and van den Berg (1974) uses bedding as the reference direction. The basic assumption in this method is that the undeformed particles had a symmetric distribution about the bedding trace. In fact this consideration, based on a theoretical analysis of sedimentary fabrics by Dunnet and Siddans (1971), led to the choice of bedding as the reference.

The third strain analysis method available for use with complex initial markers, such as sedimentary particles, is from the work of Elliott (1970). The mechanics involved in plotting distributions results in the pattern for the deformed state being very similar to that of the pre-tectonic fabric. By studying the shape of the pattern it was claimed that the initial fabric type can be distinguished from within a large range of possibilities and with this information the position of particles which were initially circular can be located. The tectonic strain ratio for a section can, therefore, be determined from the deformed distribution alone and the method appears to be capable of universal application. The Matthews, et. al. (op. cit.) technique is restricted to fabrics symmetric about bedding and cannot be applied to cases where cleavage is parallel or perpendicular to bedding unless the fabric was initially random. Symmetric initial fabrics only can be analysed by the Dunnet and Siddans (1971) procedure as the deformed pattern is progressively unstrained until symmetry is accomplished.

7.2 i: Discussion of the Fundamental Assumptions in the Strain Analysis Methods

Every study in strain analysis of complex initial fabrics has considered the likely starting material mainly by making a theoretical extension of the voluminous but somewhat inappropriate data supplied by sedimentologists. Igneous fabrics have yet to be studied in any detail and as quartz arenite was the rock-type studied in this particular investigation only sedimentary situations will be discussed. In any development of a strain analysis method involving complex fabrics, it is essential to know the relations between particle ratio (Ri) and long axis orientation (θ) for each member of an initial distribution. Sedimentologists, whilst recognising their intimate relationship, have long studied one parameter or the other and only on rare occasions both (Liboriussen, 1975). Even then the necessary information linking the two parameters for specific objects was lacking. As a result, theoretical delineations were made of initial sedimentary fabrics by Elliott (1970) and Dunnet and Siddans (1971), which, because of the imprecise basic data, must be open to doubt. In all the methods mentioned errors are introduced if the actual initial distributions depart from the expected patterns. An analysis of unstrained sedimentary material was, therefore, considered *vitally* important to provide at least some definitive information of the type suitable for the strain methods that had evolved. Testing the methods of Matthews, et. al., and Dunnet and Siddans would first involve the measurement of Ri/θ factors for a wide diversity of initial fabrics in several planes viz. perpendicular to S_0, parallel to S_0 and 45° to S_0. Each population on a plane could then be transformed into the strained state using equations presented by Elliott (1970, p. 2235), for a homogeneous pure shear whilst employing several cleavage/bedding angles. Several strain states for each orientation could be produced and the resulting data analysed by the computer programs *Strane* (Dunnet and Siddans, 1971) and *X Rot* (Matthews, et. al., 1974). The finite strain values (R_s) as determined by the programs could then be compared to actual strain values used to produce the deformed fabrics. This procedure should provide a sufficiently rigorous test of the usefulness of these methods. With such data all methods could be tested to determine their usefulness. However, it was decided to restrict the program to the most generally applicable method i.e. the Elliott method. From a consideration of the literature this method had the potential to deal with many initial fabrics and was not limited by factors such as the angle of the cleavage and bedding.

7.2 ii: The Elliott Method

After the merits and disadvantages of the various techniques had been investigated from published work and the Elliott method chosen, the next step was to measure and characterise some initial fabrics. Suitable material was not always available and much more work is required to properly characterise even the types of sedimentary environments chosen in this project. Some of the oolitic material came from regions that had undergone tectonic activity but in each case translation appeared to be the major deformation mechanism. Ideally, material should be taken from thin, flat-lying cover sequences protected from orogenic activity. The clastic rocks used were very close to this situation, but attempts made to obtain more material were unsuccessful. Each section measured was plotted using the polar graph method of Elliott which plots each object as a point. The radial distance from the centre is equivalent to

$$\tfrac{1}{2}\ (\ln (= \text{log natural})\ (\frac{\text{long axis}}{\text{short axis}}) = \varepsilon$$

The polar angle is twice the angle (θ or ϕ) between the object long axis and the reference line with a positive and negative notation as shown on Figure 7.1. Distributions are normally contoured by the Mellis method using a counting circle size which best brings out the shape of the pattern.

To understand the relevance of the measured sedimentary fabrics to the technique, the form of various distributions envisaged by Elliott must be studied as well as the way in which they were applied. An undeformed distribution will lie close to the origin of the polar graph and the effect of strain is to move the plot to the right. Transformation equations relating Ri/θ values to Rf/θ values have been plotted on a Shape Factor Grid by Elliott (1970). The contoured shape of a particular distribution changes little during strain until quite high values are reached. The initial patterns described by Elliott that are most likely to be encountered in nature are listed below.

i Circular contours centred at the origin representing objects of random orientation with various axial ratios. They are modified to ellipses by strain.

ii Deltas and hearts. Unimodal fabrics with thin ellipses most strongly aligned and coincident with the bedding trace. The latter trace being a line of symmetry.

iii Cigars and ovals. Bimodal fabrics with 90° between the modes. A banana shape reflects a bimodal fabric, with less than 90° between the modes. All these three patterns are centred at the origin.

When deformed, the strain is taken as being the centre of the distribution for groupings i and iii. In the case of a delta or heart the base line is determined and its intersection with the line of symmetry is claimed to give the location of points that were initially circular and, hence, the tectonic strain ratio. When the distribution is undeformed, the unstrained bedding-trace should, according to Elliott, closely correlate with the symmetry line of the delta or heart.

The Elliott method commonly uses the cleavage trace as the reference line, that is, the cleavage/particle long axis orientation angle is ϕ. On the cleavage face (XY), the stretching lineation, X, is used as the reference. Elliott, though, had such confidence in his method that he determined an initial circle point (ICP) on XY that was some way off the observed lineation and proceeded to use the line joining this point and the origin as X. In a section at right angles to cleavage, the cleavage trace *must* be taken as the maximum elongation direction following the commonly accepted relation of maximum shortening being perpendicular to a 'slaty' type cleavage (Borradaile, 1974b). If this assumption holds generally then, during homogeneous pure shear of a sedimentary fabric, any particles that were originally at the origin (i.e. circular) will be transformed into ellipses whose long axes coincide with the cleavage trace (or X on XY). Therefore in analysing strain by the Elliott method ICPs should be looked for along the principal direction of the section involved. If cleavage is not the XY plane of the finite strain ellipsoid (Etheridge and Lee, 1975) then initially circular points would lie off the cleavage (or mineral lineation on XY) direction but all the evidence from the simplest, most regular initial objects (point source reduction spots) does not support this (Wood, 1971). P.F. Williams (1976) has indicated that for cleavages defined by markers where there was an initial preferred orientation, the resulting cleavage will be at an angle to XY. The considered situation was, however, defining cleavage as the maximum preferred orientation of markers which is generally not the case. A low degree of initial preferred orientation will give a low departure of cleavage and XY which will be further reduced during strain by convergence.

7.3: Sedimentary Distributions on Elliott Plots

Unstrained oolitic and oncolitic limestone, quartz arenite, and litharenite were studied in this pilot project.

Figure 7.2(i), a, c, and d (also see Figure 7.3 c and d) are of a quartz arenite part of the Viking Sandstone, Joffre oil field, Alberta (see Zimmerle and Bonham, 1962). This formation is within a gently dipping cover sequence resting on the craton and to the east of the Rocky Mountain Thrust Belt. A good approximation to a delta distribution is seen in Figure 7.2(i) a, taken at right angles to the bedding whose trace on this section is the reference line. The two contour is shown in preference to the three contour which had the same overall shape but was much less continuous. This pattern is unimodal (Elliott, 1970, p.2224) which, because of the double angle nature of the plot, has a dispersion of approximately 45° on either side of the line of symmetry itself + 25° (+ 50° on an Elliott Plot) from the bedding trace. The mid-point of the base is not at the centre of the graph but is 0.04 ε units away on a line -15° (-30° on an Elliott Plot). These values could vary as the irregularities of the distribution do allow somewhat differing interpretations. An imbricate structure is demonstrated by the inclination of the longest and thinnest grains to the bedding direction. For most of the rock types discussed in this section, the Ri/θ plot of Dunnet (1969) has been employed which in the case of bedding perpendicular sections gives a measure of the degree of imbrication (Figure 7.2(ii)). The Ri/θ plots do not clearly demonstrate the fabric involved and the contoured polar plots are preferred.

When sections are cut parallel to the bedding in the Viking Sandstone material (Figure 7.2(i)c) the one contour of the Elliott Plot gives a roughly circular or random distribution (Figure 7.2(i) c, d) though the three contour shows more ordering. The band of higher concentration running through the centre of Figure 7.2(i)c represents two modes at 90° to one another, probably representing long axes of grains that lay along the current flow and at right angles to it. Because small scale palaeocurrent indicators show high degrees of variance, this type of time-consuming analysis will have little application from this point of view. The contrasts between Figures 7.2(i), c and d,do, however, illustrate significant variation in the pattern in a short distance as both sets of measurements were taken from the same polished block.

Figure 7.2(i) b is of the Permian Penrith Sandstone of England, an aeolian dune sand with a demonstrable lack of features related to "post-depositional (tectonic) compression" (Waugh 1970, p. 1235). The particular specimen measured was cemented with silica, the extremely well rounded grains being clearly delineated by iron oxide dust trails (Figures 7.3 e and 7.4 a). The section was cut perpendicular to bedding. The distribution in Figure 7.2(i) b is perhaps best described as a rectangle whose most sensible line of symmetry would be at $+22\frac{1}{2}^{o}$ from the bedding trace. Again the line of symmetry intersects the base a short distance from the origin. Figure 7.2(i) b would appear to be very similar to Figure 7.2(i) a but here the apex of the delta has been truncated due to a lack of long, thin grains eliminated by the more effective attrition of wind action.

Specimen 41834 is a litharenite from the Permian Risdon Sandstone of southern Tasmania. These sedimentary rocks form a thin generally near horizontal cover sequence which has only been involved in epeirogenic activity since deposition (Spry and Banks, 1962). It is a reworked glacial deposit containing quite angular grains which range considerably in sphericity and extreme shapes are present (Figures 7.3 f and 7.4 b). Fortunately, the original detrital grains had been coated with dust prior to diagenesis because there is evidence for considerable particle shape modification by silica overgrowths (Ellis, 1974). Only the outlines of the original grains were measured in this study. Diagenetic quartz overgrowths can considerably reduce particle axial ratios if applied evenly over the surface of the original grain. Such activity modifies post depositional fabrics and throughout this excercise only grains, whose prediagenetic outlines were obvious, were studied (see Gibbons, 1972). Hence shape factors considered were pre-diagenetic; orientation factors reflected depositional conditions as well as compaction. The Risdon Sandstone plot has no readily discernible characteristic shape when the two contour of a small counting circle is used (Figure 7.5 a). 20% of the grains have axial ratios greater than 2/1 and these have a tendency to be within $\pm 30^{o}$ of bedding (Figure 7.8(ii) a) but marked preferred orientation of the inequant grains is lacking. 45% of the grains have axial ratios less than 1.5/1 and the overall distribution is much less compact than any of the arenites discussed previously. In terms of dispersion of pattern this sample represents nearly the opposite end of the spectrum to the aeolian Penrith specimen.

Figure 7.6 a and c are of the Carboniferous Lion Creek Limestone, Queensland (Maxwell 1960, p. 174) and figure 7.6 b is an oolitic limestone from the Northern Calcareous Alps, Upper Bavaria (Fabricus 1967, plate 1b). The Lion Creek Limestone is known to have undergone body rotation whereas in the Northern Calcareous Alps translation by over-thrusting is important (Oxburgh 1968, p. 18), but in both cases microtextural structures indicative of *strain* are absent in thin section. Only the true ooliths were measured in Figures 7.6 a and 7.6 b whereas Figure 7.6 c gives the distribution in the Lion Creek limestone of the superficial ooliths i.e. those grains with large nuclei and few coatings.

Figures 7.7(i), 7.7(ii) and 7.8(i) are of Jurassic rocks from the Cotteswold Hills, England where the regional dip is less than 1° and tectonic influence is negligible (Ager, 1956). Figure 7.7(i) records the distribution of ooliths of various sizes within cross-bedded limestone. Figure 7.8(i) shows the shape factor derived from small oncolites (pisoliths) in the Pea Grit from Leckhampton, England.

For all the oolitic samples, the distributions are much more compact than those for the samples of quartz arenite described above. Measurements on unstrained chamositic oolites by Badoux (1970) suggested this would be the case but unfortunately in his study axial ratio and orientation information was not linked definitively as in the Elliott Plot. In the oolite specimens of Figure 7.6 a and b there is a tendency towards a heart or delta pattern, but, because of the position of the centre of the graph, these distributions represent *bimodal* and not unimodal fabrics. There is approximately 50° between the modes in Figures 7.6 a and b. If these distributions were to be considered unimodal then the intersection of the line of symmetry with the base of the pattern would be an appreciable distance from the graph centre. The plot of the superficial ooliths (Figure 7.6 c) from the same specimen as Figure 7.6 a gives an extremely well spread distribution that is close to circular when the 1 contour is examined. There is obviously a wide range in particle axial ratios with a good representation of long thin grains, but they are the closest to a random distribution yet found. This is puzzling in view of most theoretical considerations on factors that would control preferred orientation in post-compaction fabrics. A complete three-dimensional analysis may resolve part of the problem. The readings obtained from ooliths in the Lion Creek Limestone specimen were plotted in three groups, each containing approximately 50 ooliths. All readings came from within one 2 x 5 cm area (Figure 7.9). For each group the 3 contour was compared to that for the overall sample and significant shape variations were discovered. The more widely scattered points from this specimen in the majority would represent high and low cuts of superficial ooliths which the high contour density would ignore.

The very compact distributions from the Cotteswold Jurassic oolite are of sections at right angles to bedding (Figure 7.7(i)). Figure 7.7(i) b shows the pattern for large ooliths from down the dip of foresets in a one metre thick cross-bedded unit. This orientation was taken in the hope that the most marked preferred orientation would be recorded. However, both plots do not have any distinctive shape that would characterise them as bimodal, random, etc. Quite a contrast is found in the one thin section between large (mean size approx. 1 mm) and small (mean size approx. 0.5 mm) ooliths which occur in discrete beds. The small ooliths have a wider range of axial ratios and the resultant Elliott plot is much more asymmetric (Figures 7.7(i) c and 7.7(ii) b). The centre of the distribution is, therefore, some distance from the origin of the graph. Despite the more noticeable alignment of the grains with the highest axial ratios a distinctive shape is still lacking in the plot.

In sections at right angles to bedding the small oncoids from the Leckhampton specimens displayed a wide range of particle axial ratios. These oncoids have nucleated on plate like pieces of fossil shell debris which would account for initial high ratios. Specimen 44390 when cut perpendicular to bedding (Figure 7.8(ii) f) had only 15% of the measured grains with ratios less than 1.5/1, whilst the proportion reaches 36% in a section parallel to the sedimentary layering. In all three sections patterns are very nondescript with the usually employed counting circle size (approx. $\varepsilon = 0.10$), though the 1 contour at this size forms a rough delta with a symmetry line 25° from the bedding trace. For each section a delta shape is produced using the 2 contour at a counting circle size of $\varepsilon = 0.24$. This is despite the very real differences between the positions of the concentrations and the different relations to bedding. The cut number 3, perpendicular to bedding (Figure 7.8(i) f and g), would appear to be a unimodal fabric with the majority of the grains falling between 2/1 and 3/1 axial ratios. The half way point along the arc defined by the small 2 contour is 10 to 15° away from bedding and the total fabric appears imbricate. The most marked divergence between bedding trace and the mean particle orientation occurs on cut 2 (Figure 7.8(i) c, d, e) where something like 30° is involved. The R_i/θ plots of Figure 7.8(ii) show the amount of imbrication quite clearly.

Figure 7.10 shows some of the range of three-dimensional shapes of the small oncoids. The bottom row of this figure illustrates the most common type of shape which is an elongate disc like form. A measurement of one of these pisoliths (bottom row far left) gives X : Y : Z ratios of 3.46 to 2.31 to 1.00. The range of shapes illustrates the difficulties of taking three perpendicular cuts and combining them to produce an average three-dimensional shape orientation factor. Observations have indicated that oncoids nucleating on more equidimensional particles such as crinoid ossicles would themselves show a much lower range of axial ratios and, therefore, quite a different fabric.

7.4: Problems of Strain Analysis with the Elliott Method

Elliott (op. cit.) in discussing the various possible initial fabrics notes that significant errors would arise if the initial distribution were incorrectly identified, though theoretical considerations led him to believe that initial fabrics would retain distinguishing features even to high strains. Investigation of this factor is the main topic of the next section and there appears to be three main points involved.

Firstly, the actual measurements to date have tended to give somewhat irregular distributions (Elliott, 1970; Geiser, 1974; Etheridge and Lee, 1974; Mukhopadhyay, 1973; Figures 7.2, 5, 6, 7, 8 of this chapter) and are not as well ordered as the theoretically derived cases. Irregularities of this nature may be evened out by taking larger samples but they are in many cases an inherent feature of the sensitive Elliott Plot. Significant irregularities obviously will cause difficulties in the choice of lines of symmetry and position of the ICP and may lead to a large error. The tendency for ICP's to lie part way inside the distributions (Figures 7. 2, 5, 6, 7, 8) will also increase inaccuracies because they have been assumed to lie at the edges or the centre. Low variation in shape parameters will give very compact distributions with 75 measurements whereas quartz grains in an extremely well rounded and sorted quartz arenite may require 100-150 measurements to give a good impression of its distribution. More variable initial fabrics may require a very high number of readings to show a good recognisable distribution, if one exists.

A second consideration follows from theoretical analysis because it is well known that particles with axial ratio below 2:1 have much more variable long axis orientations that more inequant grains (Griffiths, 1967). Another observation is that most sedimentary particles lie in the axial ratio range of between 2 to 1 and 1.3 to 1 (Moss, 1966). Hence we would expect that the definitive grains of say a delta, the thinnest and longest, would be poorly represented (Figure 7.2(i) b). Likewise the extremities of the arms of a banana shape may be expected to be quite short and blurred with the grains about the ICP for this reason.

The third and most important factor concerning the distinctive nature of initial patterns is that a general delta shape has been shown to represent both a unimodal and a bimodal sedimentary fabric with very significantly different positioning of the ICP's (compare Figure 7.2(i) a with 7.6 c). Figure 7.11 b was constructed by deforming, homogeneously, the distribution of Figure 7.6 b with the initial centre of the graph moving along the reference axis. This figure illustrates the type of errors involved by taking the incorrect step at this point in the procedure. If the three dimensional check of internal consistency is applied ($^{X}/Y \cdot {}^{Y}/Z = {}^{X}/Z$) then an error may be immediately detected. However it may be possible that errors in the other two principal sections may be approximately balanced or as in Figure 7.11 the two ICP's from very different initial fabrics may lie at about the same radial distance from the centre of the graph, i.e. the finite strain in both cases is similar. Perhaps more significant is the angular error involved of about 30° if the distribution in Figure 7.11 truly belongs to a bimodal pattern but is taken as being unimodal. If the analysis is being made on the XY section so as to determine the X direction (see method in Elliott, 1970, p. 2231) then this angular discrepancy will lead to a rotation of the reference line and the remaining two sections cut as principal planes on this basis will in fact be at high angles to the actual YZ and XZ planes of the finite strain ellipsoid. Another approach to apparent angular error through lack of knowledge of the initial pattern is to consider that the deformation fabric (cleavage) is not directly relateable to the finite strain in the rock. Because of the complexity of strain analysis using variably shaped and oriented particles it seems difficult to be precise about relationships between finite strain and cleavage. The angular differences discussed by Etheridge and Lee (1974) are just as likely to be a result of poor knowledge of the pre-strain distribution of their strain indicators than to a lack of correspondence between directions of maximum shortening and cleavage.

7.5: Modifications to the Elliott Technique

In view of the fairly serious difficulties presented by the investigation of actual initial fabrics, the method of strain analysis proposed by Elliott was modified. The most significant change was to assess the position of the ICP by fitting comparable unstrained fabric plots to the deformed plots. This replaced the finding of ICPs using intersections of base lines and lines of symmetry for unimodal fabrics, centres of distributions for random fabrics, etc.

For sections at right angles to cleavage, the cleavage trace was always taken to contain the ICP. Besides fitting undeformed and deformed plots by shape, the bedding trace relationship would have to be taken into account and also the bedding/section dihedral angle. Sedimentary fabrics vary considerably from sections perpendicular, to those parallel, to bedding and, in low to moderate strains, sections in the deformed state that are at a high or low angle to bedding would have held a similar relation prior to deformation. This factor can be taken into account when assessing which initial fabric plot is to be used. The initial fabric investigation was only concerned with sections either parallel or perpendicular to bedding. Strongly imbricate fabrics would have different fabrics at 45° to bedding and more complete analysis is warranted as intermediate situations are likely to be found in practice.

7.6: A Pilot Project to Test the Modified Elliott Approach - Strain Analysis of a Cleaved Oosparite Affected by One Deformation.

Before this modified approach was applied to quartz arenite of the Frankland Range, it was decided to test the applicability of the method to a simpler situation. The folded and cleaved Ordovician Gordon Limestone at Mayberry was chosen because of the comparative simplicity of the structural style and the fairly regular, compact distributions of the oolith strain indicators. The scope for error in the determination of strain is lessened if the initial markers do not have a wide range of initial axial ratios because the length of the plot intercept on the reference direction is reduced. The ICP would normally be within the plot or at its margin, very rarely outside.

The folding and cleavage at Mayberry is the result of the mid-Devonian Tabberabberan Orogeny which though polyphase essentially gave rise to a dominant cleavage during one episode. The cleavage is everywhere steeply dipping and the folds are for the greater part gently to non-plunging. All the specimens for this strain analysis were collected from Grunter Hill after discussions and field visits with Mr. B. Pierson who was conducting a sedimentological study of the Gordon Limestone in the Mayberry district.

For all the Grunter Hill specimens the direction of the cleavage plane was marked and then cut with a diamond saw. Some difficulties were experienced in specimens that were later found to have less than 25% shortening across the cleavage because the cleavage was poorly defined. In most cases the X lineation on XY was quite clear though errors would be expected towards the lower strain end of the spectrum. It appears that when most of the initial markers have axial ratios less than 1.3 to 1 then an X lineation can accurately be placed at $R_{XY} > 1.50$ to 1. A possible source of error lies in the possibility of some initial fabrics giving rise to high preferred orientations of object long axes (elongation lineation of Rast, 1966, p.25) at some angle to the maximum elongation direction. Sections YZ and XZ were cut at right angles to XY and oriented thin sections prepared for the three principal planes. Because initial fabric variations are expected from one sedimentary layer to another all three sections should be taken from the same layer and be as close together as possible. Negative prints were prepared by placing the thin sections directly in an enlarger. Oolith axial ratios and orientations were then measured using X on XY and the trace of XY on YZ and XZ as reference directions. Cleavage was assumed to parallel the XY principal plane of the strain ellipsoid. The standard sample was 75 and, whilst 50 seems suitable for the Rf/ϕ graphs of Dunnet (1969), the larger number may even be inappropriate for the polar graph (see shape variations for different samples with specimen 35872, Figure 7.9). The fold in the Gordon Limestone at Grunter Hill is open and nearly upright and the X lineation is down the dip of the cleavage plane. This geometry gives a high dihedral angle between bedding and both the XZ and XY sections. Bedding and YZ have a low angular relationship.

The resulting Elliott plots show no distinctive heart, delta, oval, cigar or banana shapes and to proceed on the lines originally suggested for this technique would be difficult. Figure 7.12 AYZ, XZ, BXZ, DXZ, EYZ are close to symmetric but their patterns are not particularly oval and would not seem to have been random initially. Some shapes e.g. DXZ, EYZ are quite similar to the truncated deltas of the Penrith sandstone and is perhaps an expected pattern in oolite with few inequant grains to define the bedding direction. In these two sections the position of the deformed bedding trace also corresponds to this initial fabric, suggesting the ICP lies to the edge of the distribution furthest from the origin. Several plots are markedly asymmetric with respect to the cleavage trace and hence any strain analysis technique which employed the assumption of initial randomness would be expected to fail in these cases. To test the errors involved in such an approach the strain was estimated by taking the centres of each distribution. By taking a common assumption about the nature of all the initial fabrics, the pitfalls of *forcing* the R_{XY}, R_{YZ} and R_{XZ} values into conformity are avoided. Table 7.2 gives the values for XY, YZ and XZ and then compares $\varepsilon_{XY} + \varepsilon_{YZ}$ with the measured ε_{XZ}. 44308 and 43310 (Figure 7.12 a, c) are internally compatible but the other three plots show quite high departures. An inspection of the Elliott plots shows a good correlation between the uncommon shape factors and the incompatible results from Table 7.2. Clearly if only two principal sections (or three cuts with no way of making an independent check) had been measured then the scope for error is considerable even with ooliths

TABLE 7.2

	ε_{XY}	ε_{YZ}	$\varepsilon_{XZ} = \alpha$	$\varepsilon_{XY} + \varepsilon_{YZ} = \beta$	$\frac{\alpha-\beta}{\alpha}\%$
A 43308	.274	.178	.471	.452	+4
B 43309	.300	.241	.434	.541	-25
C 43310	.227	.187	.398	.414	-4
D 43311	.208	.247	.345	.455	-32
E 43312	.194	.092	.338	.286	+15

Centres of distributions taken as the ICP i.e. assume the fabrics to have all been initially random.

whose initial variability is quite low. From Table 7.2 the three-dimensional strain values, using the initial random assumption, all lie in the constriction field of a log/log deformation plot (Wood, 1974), except for 43311 which falls in the flattening field. Considering the lithological similarity of the specimens and the restricted area over which they were collected it is to be expected that they would all have the same type of tectonic strain. The difference between 43311 and the other four samples would suggest an error in the strain assessment which further reflects upon the method employed.

In applying the modified technique, shape factors from unstrained specimens were fitted to the plots from the deformed specimens. The analysis was somewhat limited by the lack of more extensive initial data particularly the lack of data from sections at varying angles to bedding. For the deformed material, many sections gave two or even three good shape fits with different values for the strain. These possible values were often of the same order of magnitude and rarely

TABLE 7.3

		R_{XY}	ε_{XY}	R_{YZ}	ε_{YZ}	R_{XZ}	ε_{XZ}	X	Y	Z
A	A	1.81	.297	1.55	.219	2.76	.508	+71	-5	-39
43308	A'	1.64	.247	1.43	.179	2.33	.423	+57	-4	-33
B 43309		1.71	.268	1.51	.206	2.56	.470	+64	-4	-37
C 43310		1.57	.226	1.48	.196	2.34	.425	+54	-2	-34
D 43311		1.52	.209	1.41	.172	2.14	.380	+48	-3	-31
E	E	1.55	.219	1.29	.127	1.99	.344	+46	-6	-28
43312	E'	1.46	.194	1.20	.092	1.97	.338	+37	-6	-21

were good shape fits found such that the ICPs lay at either extremity of the plot. Further selection of the strain ratio for each section took into account the minimisation of the angular distance of the proposed ICPs from the reference line, the compatibility of the product of R_{XY} and R_{YZ} compared with R_{XZ} and the bedding/section dihedral angle compared with that of the unstrained plot. By taking these factors into account the three-dimensional strain states were determined (see Table 7.3 and Figure 7.13) and all plot in the constriction field. Specimens 43309 and 43311 perhaps best show the value of the Elliott method as it detected considerable departures of pattern for their XZ and YZ sections respectively. For these samples the strain values on the other sections were quite closely defined and it was found that, for internally consistent results to be obtained, the ICPs for the elongate plots would have to lie at the point of highest concentration of ooliths. This agrees with the data from Figure 7.6 a where the origin was closely associated with highest density of points. The patterns of 43309 XZ and 43311 YZ were quite unlike any measured from unstrained rocks and a further complicating factor is that the two sections with similar patterns have very different bedding/cleavage dihedral angles.

7.6: i) <u>Discussion of the constrictional tectonic strain values of the oolite markers.</u> Most cleaved rocks whose strain states have been investigated to date have plotted in the flattening field of the deformation plot. The cleaved Grunter Hill oosparites lie in the constriction field possibly due to the influence of large ductility contrasts between the sparry nuclei and micritic coatings of individual ooliths (Tan, 1974). Most of the rock is sparry calcite and the layered coatings may only form approximately 20% of the total (Figure 7.14 a). Following the work of Gay (1968), the high volume concentration of coarse calcite in all the specimens would indicate that the three-dimensional strain state from the oolith markers must be close to the bulk tectonic three-dimensional strain of the rock. It is therefore expected that the true strain state for the Grunter Hill specimens would be close to the 'plane strain' line but certainly not far into the flattening field. An oncosparite from the same area plots very close to the boundary between the two fields (Seymour, 1975) in a situation of lower ductility contrast between markers and matrix. The work presented in this section was expanded in an Honours thesis by D. Seymour (1975) under the supervision of the writer.

7.6: ii) <u>Conclusions from the application of the Elliott Method.</u> The chosen approach would always supply several different shape fits leading to more than one possible strain ratio for any section. A further complication is added by the difference in pattern between the same distribution as viewed from two sides. If a specimen is cut as a block, then the shape factor viewed from the outside looking in, on one section, is reversed and rotated through 180° if plotted as viewed from the inside looking out. Reversals may well increase the number of ICPs given by this modified Elliott method.

In order to reduce the number of choices per section to one, several other factors have to be taken into account. It seems essential to have some form of internal check on the validity of the results to allow spurious results to be rejected. This is most readily achieved by sectioning along the *three* principal planes of the strain ellipsoid but there is some danger that inaccurate results may be forced into conformity. In suitable sections, such as those at high angle to bedding, the position of the bedding trace should eliminate some of the proposed shape fits and should always be checked.

Despite some uncertainties the Elliott method demonstrated its versatility in being able to clearly indicate differences in shape factors which are important when considering initial fabrics. Rather than allowing the operator to carry, throughout a strain analysis project, only one or two basic initial fabric assumptions, this method requires consideration of many variations. The investigation of initial distributions revealed complex differences even between closely associated sections and these should also be present in the deformed situations. Employing the Elliott approach on deformed oosparites has given good results both from the general uniformity of the strain ellipsoid type over a small area (Y shortened between 2 and 6% for five lithologically similar specimens) and the three-dimensional compatibility.

7.7: Strain Analysis within the Frankland Range Quartz Arenite

7.7: i) Structural position of the specimens. A collection was made across the major discontinuity of the eastern Frankland Range starting on the right way up, moderately dipping limb of the large close fold (see profile 7.15, specimen 44377). Still on this limb but within metres of the metamorphic slide specimens 44374 and 44370 were taken. Soon after the change to bedding-parallel cleavage, specimen 44371 was collected but proved to be too well recrystallised for strain analysis. Within this upper structural unit of major isoclinal folds, 44375 was studied to assess the geometry and orientation of the strain ellipsoid in a situation of cleavage and bedding being parallel. 44372 was from the closure of a D_1 asymmetric coupled fold and 44375 was on its external limb (Figure 7.14 b).

Specimens 44366/67 are from a very similar structural situation to the overturned, steeply dipping limb of the close fold mentioned above (see profile 7.15). At this locality S_1 dips at between 15° and 150°, though generally at low angles, and occasionally another cleavage is measurable in the field in a similar orientation. 44367 is an example of this localised cleavage whereas 44366 appeared to represent the more general, low-angle, S_1. On the right way up limb of this easternmost set of folds several specimens were collected from closures of parasitic D_1 folds to determine the amount of deformation associated with these folds and the relations between strain and fold geometry. 44373 is from a minor fold at the tightest part of the major fold (Figure 7.16 a) which does become more open towards the zone where 44369 and 44368 were collected (see Figure 7.16 b for the style of minor folds in this area).

7.7: ii) Working methods and practical problems. Initially all specimens were sectioned parallel to the cleavage and a thin section cut either parallel or perpendicular to the cleavage. The latter was the case before the strain analysis project and the previously oriented section was used in the measurement of one group of specimens in the project. Valuable information was gained from these sections as to the suitability of the material for further work. Many samples were rejected because the strain indicators were not clearly outlined. Recrystallisation was often responsible for the erasing of the dust inclusion trails around detrital grains, which in numerous low strain cases were equally indistinct due to their original poor development. Well developed microscopic effects of later deformation events also meant the rejection of specimens. When S_1 was the only surface present pressure solution was invariably in evidence but this rarely was of sufficient strength to require the discarding of the specimen. Later cleavages varied in their intensity from slightly modifying most grains in the slide to being spaced zones of pressure solution with little effect in the intervening zones. High D_1 strain states gave rise to the best fabrics and these were the most readily acted upon by the later deformations to further complicate the analysis. A very high proportion of the material collected was rejected because of these difficulties and this severely limited the planned program of investigating strain variation around small scale folds. Extensive collecting at the most favourable sites would be required to overcome this problem.

Various techniques of measuring the Rf/ϕ data were employed. Most commonly thin section photomicrographs were enlarged or for clearly defined, large quartz grains a negative print was made by placing the thin section directly into an enlarger. Difficult specimens with recrystallisation were measured directly on a microscope stage or by a combination of negative print and a projected image, in analysed light, on a ground glass screen. Examples of the material studied are presented in Figures 7.17, 18, 19, 20.

7.7: iii) Elliott plots and the assessment of strain. The first approach adopted, involved cutting in the cleavage direction and to assume that it represented the XY plane of the strain ellipsoid.

Oriented XY thin sections were used to determine the X direction which did, however, vary in definition. If the mean of the mica beards was not a clear concentration of measurements then the accuracy of the X orientation was open to question. This was the case with specimen 44378 and in its immediate neighbour, 44377, X was not too sharply defined. Problems of this nature were the prime cause for the adoption of the method suggested by Roberts and Siddans (1971) for a later group of specimens.

Once X was determined thin sections were cut in the YZ and XZ directions attempting to keep the sections as close together as possible whilst sampling the same sedimentary layer. Some specimens consisted of alternations of silt and sand sized grains which deformed somewhat heterogeneously also giving rise to slight cleavage refraction. The degree of heterogeneity brought about by lithological variations may be quantified in some specimens but was not attempted in this study.

Polar plots were prepared for each section and contoured, generally using a counting circle diameter of about $\varepsilon = 0.100$. Several combinations of densities and circle sizes were used to study the way these variations affected the shape of particular patterns (Figure 7.21). Counting circles around $\varepsilon = 0.250$ gave a much smoother outline, particularly for the 1 countour, but when applied alone to deformed examples produced more possible ICPs. By constructing a 3 or 2 contour at the $\varepsilon = 0.100$ size, in conjunction with an $\varepsilon = 0.200$ 1 contour, reasonably restricted fits could be found. In general though, the best operating size appeared to be about $\varepsilon = 0.100$ with a 2 contour giving clearer patterns than the 3. The larger counting circle does perhaps tend to take an overview of the whole distribution and hence approaches the Dunnet and Siddans (1971) method which considers the symmetry of the whole plot. The comparison made here is not strictly applicable because the Dunnet and Siddans method of assessing symmetry takes much more notice of concentrations within the distribution and not just of the outermost members.

Observations on the specimens with negligible recrystallisation indicate that, in the unstrained state, the quartz grains were extremely well rounded and quite spherical (Figure 7.22 b). Qualitatively, they in fact most closely approximate the aeolian Penrith Sandstone example and possibly lie somewhere between that and the type of material represented by the Viking Sandstone. The general size of the Elliott plots for the deformed material is within the range found for undeformed sandstone and many distributions approximated deltas (Figure 7.23, AYZ, XZ, BXY, YZ, XZ, CYZ, XZ, DXY). As might be expected with well abraded quartz grains, few inequant grains were present originally so the delta shapes are often squat and not clearly defined. To claim that all of the deltas represent initial unimodal fabrics is likely to be incorrect especially when the angular difference between bedding traces and delta lines of symmetry is considered (Figure 7.23 BYZ, XZ). Several shape factors are very close to the 'truncated delta' pattern of the Penrith Sandstone plot and, solely regarding shape, virtually all the patterns are within the types determined in the limited examination of undeformed fabrics.

Anomalous results come to light when the dihedral angle between principal planes and bedding is taken into account. Figures 7.23 BXY, CXY, DXY have quite good delta shapes and, in the latter two plots, the bedding trace is in the correct position for an imbricate initial fabric. All these sections are, however, at low angles to bedding and if the data from the Viking Sandstone is typical then an elliptical pattern with a cigar shaped concentration is to be expected. It may be possible to have unimodal fabrics in the plane of imbrication but more extensive studies of undeformed quartz arenite are required before the three-dimensional fabric can be characterised. Within the one specimen the patterns from two sections at very different angles to bedding can be extremely similar (Figure 7.23 B,XY and YZ; E,XY and XZ). Such an unexpected situation is difficult to explain on the basis of the presently available data from undeformed specimens. As with the oosparites from Grunter Hill the above observations seem to indicate the complexities of sedimentary fabrics and the dangers of making simple assumptions that cover an intricately variable reality.

Some attempt was made to see how well the assumption of a near random initial fabric would fit data from the quartz arenite. Specimen 44374 (Figure 7.23 a) is taken as a typical example which illustrates the inappropriateness of this type of fit for many of the sections. If a random fabric is deformed the ICP should move along the principal direction in any section and this line should be the long axis of the resulting elliptical plot. For 44374 YZ (Figure 7.23) the best fitting ellipse has a long axis at $+10\frac{1}{2}^{\circ}$ to the cleavage trace on the section whilst XZ is closer at $2\frac{1}{2}^{\circ}$. By taking the three ellipse centres there is also a significant difference between the measured R_{XZ} and the calculated value. Similar asymmetric specimens from Figure 7.23 would also show problems of this nature; AXY was, however, a good fit to an ellipse, the long axis coinciding with a moderately developed beard alignment in thin section. This relationship in fact fixed the location of X and hence YZ and XZ. As AXY was at a low angle to S_0 the near random initial fabric accords with the work on undeformed rocks.

By particularly noting the bedding/section dihedral angle, the appropriate sections from the Viking and Penrith sandstone plots were compared with the data from the deformed quartzite. The goodness of shape and bedding trace fits was assessed to determine which of the measured initial fabrics most closely resembled the deformed fabric being studied. The best fits were tested by the internal check and if the difference found to be high then further fits were searched for. Specimen 44375 (Figure 7.23 b) was studied using both the Penrith and Viking plots for the YZ and XZ sections. The XY section was parallel to bedding and hence the Viking bedding parallel distributions were used in the matching process. In YZ and XZ fairly good fits were obtained from both initial fabrics but the Viking values failed to satisfy the internal check. Purely using shape fit and bedding trace relationship the best Penrith results were $\varepsilon_{XZ} = 0.145$ and $\varepsilon_{YZ} = 0.197$ (Table 7.4) which clearly cannot be the actual strain if the cleavage is the XY plane of the strain ellipsoid and X was positioned correctly (in this specimen X on XY was clear in hand specimen and in thin section). By bringing the question of internal compatibility into the elimination procedure the final values were chosen (Table 7.5). It is obvious that by using this method with no internal means of checking the validity of the results then considerable errors could be involved. Good compatible fits were found for 44377 (Figure 7.23) using the Viking plots which agrees with the approximate delta patterns for CYZ and XZ. XY, at a low angle to S_0, was matched with the bedding parallel plot. YZ and XZ were matched with the bedding perpendicular plot which for XZ may produce some error as this section in the deformed state is only 60° to the bedding. $(\varepsilon_{XY} + \varepsilon_{YZ}) - \varepsilon_{XZ}$ is 0.026 for the determined strain values, possibly as a result of the uncertainty with XZ. Very good fits for 44366 were obtained using the Penrith perpendicular to bedding plot for the two sections at high angles to bedding. The XZ plot was similar to the bedding parallel Viking plot but did show some departures and, except for the bedding/section

TABLE 7.4

Strain values as ε_s

Perpendicular to S_0	XZ	Penrith	<u>0.184</u> 0.190 0.145	Viking 0.230
Perpendicular to S_0	YZ	Penrith	<u>0.126</u> 0.197	Viking 0.120
Perpendicular to S_0	XY	Viking only	<u>0.065</u>	

Underlined values taken.

Table 7.4 shows the ε_s values for various shape factor fits between the strained material and the two undeformed quartz arenites characterised.

TABLE 7.5

SPECIMEN	ε_{XY}	ε_{YZ}	ε_{XZ}
44374	0.090	0.241	0.337
44375	0.065	0.126	0.184
44377	0.093	0.198	0.265
44378	0.030		
44366	0.009	0.152	0.161

Strain measurements from the quartz arenite specimens made by following the modified Elliott method.

dihedral angle, was much closer to the Penrith bedding-perpendicular plot. The value of considering the angular relations, rather than just plot shapes and bedding traces, is amply demonstrated here.

7.7: iv) Heterogeneous deformation and deformation mechanisms in D_1

44366 was cleaved in a regular fashion at the mesoscopic scale and in thin section a very weak second deformation was found to be patchily developed even within one thin section. Of significance is the presence of zones of virtually coplanar high and low strain throughout the rock. The mesoscopic XY, YZ and XZ surfaces all contain two strain domains with a few degrees of cleavage or X lineation refraction between them (Figure 7.24). It is possible that this is the result of a superimposed deformation which would have to be of a kink like nature because of the planar boundaries between the zones which are particularly illustrated in the 'XZ' section. Compositional or grain size differences cannot be invoked and a kink style deformation is unlikely as much higher degrees of rotation are required before the rotated limb would suffer noticeable strain. The observed characteristics tend to favour the suggestion of heterogeneous deformation such that one part of a rock with fairly uniform lithology has yielded further than another. Perhaps, during the deformation of this type of material, a point is reached where the strain rate increased rapidly but in 44366 deformation was arrested at about this stage. This particular specimen should thus supply information about the progressive deformation of some quartz arenite samples and the nature of the transition from low to high strain states.

Elliott plots were prepared for the low strain regions of the mesoscopic XY, YZ and XZ sections (Figure 7.23 e) and the high strain region of 44366 (Figure 7.25 b). Very good shape comparisons were found between the three low strain sections and plots of the Permian

Penrith Sandstone material. These fits indicated that, within the low strain areas, the strain was three-dimensionally compatible. Though the Elliott plots have different patterns from the 'XZ' low and high strain areas, estimates of the differential strain can be made. R_{XZ} values are 1.36 and 2.82 respectively but the marked pressure solution effects seen in the high strain zone indicate that the strain ratio there is only an approximation. Solution would also modify grain outlines thus accounting for shape factor differences in the Elliott plots of Figure 7.25. It is apparent that the deformation mechanism had changed from the two zones as pressure solution is minimal in the lesser strain zone and quite intense in the high strain portion. Crystal-plastic processes were obviously the dominant deformation mechanism up to a strain ratio of at least 1.36 to 1 and at some point after that pressure solution became important. Whether the latter mechanism eventually took over completely or not is impossible to say but optical effects do indicate a significantly increased crystal plastic component in these zones. Also the reason for the change is uncertain as it may be related to a change of conditions (strain rate, pressure and/or temperature) or perhaps quartz under the conditions that applied during the first deformation of this area could not continue to deform by crystal-plastic means alone beyond low strains. Similar strain variations are found on 'XY' between low strain zones with restricted pressure solution to high strain and extensive pressure solution (Figure 7.24 c).

7.7: v) Exceptionally low strain states in tight D_1 folds. Immediately to the east of Murphys Bluff a prominent structural unit contains a stack of slightly verging, very tight chevron folds (Figure 4.30) which only rarely are associated with a generated axial plane cleavage. In fact the superimposed fourth cleavage is much more prominent though itself is by no means ubiquitous. A specimen (44376) was taken from the closure of one of these folds to determine the amount of strain and to see if this information could throw any light on the mechanisms of fold formation within the larger structural unit. The fold sampled is illustrated in Figure 7.22 a, which shows the paucity of superimposed effects at the outcrop scale. Thinly layered zones are openly folded by F_4 on a small scale (few cm wavelength) and the related cleavage is rarely visible. The effects of the later deformation are found only in thin section as mica films with evidence for highly localised pressure solution as their main mechanism of formation. Only extremely minor deformation at the grain scale could be attributed to the superimposed events especially in the dominant thick layers of quartz arenite.

44376 was collected from the hinge point of the fold in Figure 7.22 a, and sectioned parallel to the axial plane of the fold which was assumed to be the XY principal plane of the strain ellipsoid. Poorly developed, but consistently aligned, mica-beards in this section were regarded as the X lineation and the YZ and XZ sections cut accordingly. The polar plot of the XY section produced a fairly good delta shape especially when the symmetry line was considered in relation to the bedding trace (Figure 7.21 a, b and c). This was a good result for a section at high angles to bedding. XZ at a very low angle to bedding gave a pattern quite similar to sections of Viking Sandstone comparably oriented with respect to the bedding (Figure 7.21 e and f). Strain in both sections was exceedingly low and considering the errors in the Elliott method the two sections could not be distinguished. As this specimen was one of the first studied in the strain analysis program the data was presented using different contour densities and sizes of counting circles. Also, section XY was used to study heterogeneities brought about by size differences in the objects measured. Figures 7.21 a, b, and c show various means of presenting the plot of all the one hundred and fifty grains measured within a small area. Within 0.5 cm of this area, fifty large grains were measured and plotted (Figure 7.21 d). Differences between the patterns are considerable and, taking note of the overall very low strains, must indicate different initial shape and orientation combinations between the large and small grains. Sampling must, therefore, be sufficiently comprehensive to take an average of the material being investigated and, perhaps, the standard sample of this study (75-80) is a little low where grain size is variable.

Shape factor variations have been brought about in one sample by varying the size of the counting circle and the contour density. The delta shape of Figure 7.21 b is transformed into an ellipse by using a large counting circle and the two contour. The small circle and one contour also suggest an ellipse (Figure 7.21 a) though much more irregularly. Though the centre of the ellipse and the ICP, given by the delta plot, are quite close it seems unlikely that a section at high angles to bedding would have had a near random initial fabric. The ellipse fit is therefore inappropriate but without the section/bedding relation may have appeared the best fit. It is obvious that a few scattered outlying points may have undue influence on the assessment of strain and their importance may be difficult to analyse.

7.8: Strain Analysis in the Quartzite using three sections perpendicular to XY

After several specimens were analysed utilising three measurements on principal planes, a change to the Roberts and Siddans (1971) procedure was made. This was brought about by the occasionally poor definition of the X lineation on the cleavage (XY) plane, the overall low total strain, and the generally very low deformation in the XY plane. The XY plane was sectioned and three cuts (A'B'C') perpendicular to XY were made. A' and B' were perpendicular to each other and C' bisected the other two. From the strain ratios on these sections, the X direction can be calculated or constructed and the X:Y:Z ratios can be determined. Roberts and Siddans (op. cit.) recommend this variation where the initial fabrics appeared not be random which would be the situation for the quartz arenites discussed here. Unfortunately this approach gives a strain result for which there is no internal check unless further analysis is carried out on the uncertain assumption that the pre-deformation fabric was derived from a random one by homogeneous compaction. The variation used for random pre-deformation fabrics involves cuts A', B' and for C' to be parallel to XY. There are then two independent ways of computing λ'_z which provide the internal check. As the locus of the ICP was given by the cleavage trace on the sections A',B',C', it was at first considered that, taken with the section/bedding dihedral angles and bedding trace orientations, enough information was available to determine the strain ratios. As the strain ratios were to be decided and X positioned before the XY plane was to be studied, the change of method offered an opportunity to apply the Elliott technique without the possibility of forcing the results. If on subsequent examination of XY the X lineation could be clearly positioned then an indication of the accuracy of the strain analysis could be made.

Several specimens were rejected after the sections A',B' and C' had been cut because one or more faces displayed excessive recrystallisation or superimposed deformation. Polar plots for each face were prepared from the four specimens finally selected (Figure 7.26). Strain ratios were assessed by determining best fits to appropriate sections from undeformed arenite specimen whilst taking into account the factors discussed previously except for the three-dimensional compatibility check. If the deformed sections were either at a high or low angle to bedding it was generally seen that the range of possible shape factor fits was small (Table 7.6 (a) cut B; (b) cut B; (c) cut B' (d) cut B) or if a large range was involved then the majority of values were in close agreement (TAble 7.6 (a), cut A). Some sections in this orientation category posed problems. Section A of 44373 gave a range of values from ε = 0.054 to 0.167 with the centre of a best fitting ellipse lying on the reference line at a value of ε = 0.085. Because of the section's low inclination to S_0 the latter value must be close to the strain ratio but could not be used directly as the section was in fact 25° away from bedding. Section A of 44368 gave two quite different values with no apparent way of deciding which was the most appropriate and section A of 44372 gave no good fits despite its high angle to bedding. Significant problems were found for three of the four C cuts measured (Table 7.6 a, c, d) where a large range of values was obtained for each section and often some feature of each fit was unsatisfactory. All these sections had one factor in common; their intermediate dihedral angle with bedding, for which no comparable undeformed sections were available. Considering the

TABLE 7.6

	CUT A	CUT B	CUT C
	$A \parallel S_0$	$B \wedge S_0 = 90°$	$C_1 \wedge S_0 = 45°$
A 44367	Moderate fit to fig. 7.2d at ε = 220 ε= 0.217) Moderate fit 0.150) fig. 7.2c Good reversed fit to fig. 7.2c at ε = 0.180. ε_s = 0.185 R_s = <u>1.45</u>	Very good fits of shape + S for fig. 7.2 a, b at ε = 0.188 and 0.168 respectively. ε_s = 0.175 R_s = <u>1.42</u>	Awkward pattern ε=0.148 moderate fit to fig. 7.2 a ε=0.145 for fig. 7.2c moderate fit No fits for fig. 7.2b ε_s = 0.145 ? R_s = <u>1.34</u> ?
B 44368	$A \wedge S_0 = 70°$ Very good shape and S_0 fits for fig. 7.2a and b at virtually no strain. A good reversed fit to figure 7.2b at ε = 0.117 R_s = 1.26. ε_s = 0.00 ? R_s = <u>0.00</u> ?	1st cut $\wedge S_0 = 75°$ Good shape fit to fig 7.2a, S_0 15° off, ε = 0.106. Good shape fit and very good S_0 to fig. 7.2b, ε = 0.119 ε_s = 0.110 R_s = <u>1.25</u>	$C \wedge S_0 = 60°$ Goodish shape fit and very good S_0 to fig. 7.2b, at ε = 0.107 ε_s = 0.107 R_s = <u>1.24</u>
C 44372	$A \wedge S_0 = 80°$ ε = 0.297 good reversed shape fit to fig. 7.2a. Fig. 7.2b has no good fits. ε = 0.154 centre of distribution. No good fits yet section at 80° to S_0. ε_s = 0.160 ? R_s = <u>1.38</u> ?	$B \wedge S_0 = 10°$ Good shape fit to fig. 7.2c reversed, ε = 0.265. Good fit to fig. 7.2 d, ε = 0.262 also reversed at same value. ε_s = 0.262 R_s = <u>1.69</u>	$C \wedge S_0 = 40°$ Goodish shape and S_0 fit to fig. 7.2a at ε = .103. Good reversed fig. 7.2a fit but S_0 poor, ε = 0.194. Moderate shape but poor S_0 fit to fig. 7.2b. ε = 0.150, ε = 0.090 to 0.130 for fits of fig. 7.2c and d. ε_s = 0.130 ? R_s = <u>1.30</u>
D 44373	$A \wedge S_0 = 25°$ Two good shape fits to fig. 7.2c. ε = 0.167/0.054. Also two good shape fits to fig. 7.2d at ε = 0.118 and 0.065. Centre of distribution is ε = 0.085. ε_s = 0.100 R_s = <u>1.22</u>	$B \wedge S_0 = 80°$ Extremely good fit for shape + S_0 to fig. 7.2b at virtually no strain. Good shape moderate S_0 fit to fig. 7.2a at ε = 0.060. ε_s = 0.030 R_s = <u>1.06</u>	$C \wedge S_0 = 45°$ Good shape fit but poor S_0 to fig. 7.2a at ε = 0.190. Very good reversed fig. 7.2b shape + S_0 fit at ε = 0.155. Moderate reversed fig. 7.2a fit ε = 0.183. 0.08 very good shape fit to fig. 7.2c. ε_s = 0.150 ? R_s = <u>1.35</u>

The Table indicating the various shape factor fits used for each section and the goodness of fit of bedding trace. For each section the preferred strain ratio is given and the section/bedding dihedral angle.

differences in pattern between bedding parallel and bedding perpendicular sections it is likely that sections inclined at 45° will have different shape factors again. Because of the unsuitability of the reference plots, it is reasonable to question the validity of the strain values from the C sections being discussed. If the principal sections are used and a range of values is found for a particular section then the internal check can be used to assess which is the best of several good fits. The Roberts and Siddans technique does not allow this additional sorting criterion.

The best strain values for each section, however, were used for the calculation of the X direction and the X:Y:Z ratios. The Mohr construction, using elongations in three directions with known angular relations, was used following Ramsay (1967, p. 79-81). This method gives a simple internal check on the computation and was preferred to the analytical method of Roberts and Siddans (1971). The position of X, with respect to the deformed and undeformed construction triangle, also provides a visual check on the validity of the calculation. The pitches of the X lineation on the XY planes of the specimens from Table 7.6 are shown in Table 7.7 together with the X:Y:Z ratios. Three of the four strain states are quite similar to those found in the first group studied, in that the XY strain ratios are low. 44372 is quite exceptional with a very high ratio and there is a general tendency for the results to plot closer to or within the constriction field. Small errors in the construction probably gave rise to the unusual ratios from specimens 44368 and 44373 which are most likely examples of Y=Z. These strain states are clearly incompatible with the planar fabric seen in the field and is an indication that the problems associated with the intermediate angle sections has led to error in the determination of the three-dimensional strain.

All the evidence at this stage pointed to errors in the strain assessment for the sections at right angles to XY. The most likely source lay in the sections at intermediate angles to bedding but some other sections may be in doubt. It was, therefore, decided to cut thin sections in the XY plane to determine the orientation of X. The results are given in Table 7.7 and are compared with the results gained from the Mohr construction. Large discrepancies are obvious in each case including those where X is not located with precision and the results cast further doubt on the strain determinations for the A', B', C' sections. After X was measured, stereographic projections for the four specimens showed that in three examples cuts A' and B' were almost identical to XZ and YZ respectively (Figures 7.27 a, e; 7.28 f). The values $R_{A'}$ and $R_{B'}$ would in fact by very close to R_{XZ} and R_{YZ} which immediately required reassessment of the strain ratios previously determined for 44368 cut A' and 44372 cut A'. The latter specimen in this cut had failed to show any good fits in the initial study and a very good shape and bedding trace fit had in fact been overlooked. The strain ratio for this fit with Figure 7.2(i) a, at 1.81 was greater than $R_{B'}$ and was at least in agreement with the required relationship XZ $\geqslant$ YZ (XZ $\not<$ YZ). Following similar reasoning, the extremely good fits for 44368 section A' at no strain would have to be discarded in favour of a poorer fit (R_s 1.26) which at least allowed XZ to be slightly in excess of YZ. With no further method of internally checking the validity of the results, the strain ratios on A' and B' (for the three specimens where these approximate XZ and YZ) are similar to the results obtained from the first group studied. Very low XY ratios would be the general rule given that the $R_{A'}$ and $R_{B'}$ values are fairly accurate. Without knowledge of the strain ratio in the

TABLE 7.7

	Definition of X	Orientation of X by thin section observation. Pitch on XY	Orientation of X by Mohr construction using ratios determined by the Roberts and Siddans method. Pitch on XY	Angular difference	X:Y:Z	XY
44367	±2°	46° to NW	78½° to NW	32½°	1.53:1.31:1.00	1.17
44368	±15°	51° to NW	70° to SE	59°	1.29:0.984:1.00	1.31
44372	±5°	86° to SE	45° to NW	49°	7.07:1.25:1.00	5.66
44373	±10°	28° to WNW	63½° to WNW	35½°	1.35:0.996:1.00	1.36

X:Y:Z and XY ratios are those given by the method of Roberts and Siddans using three sections perpendicular to cleavage.

XY plane no further check can be made. If the A' and B' cuts had not approximated principal planes, the strain ratios (including C') may have been transformed into these planes once X was determined (Ramsay 1967, p.147). The large uncertainty in $R_{A'}$, $R_{B'}$ and $R_{C'}$ values did not seem to warrant this conversion in the case of 44368. The values determined gave Z greater than Y which clearly cannot be correct and the results must be discarded.

The most conclusive check on the strain determinations appeared to be the internal consistency calculation on the principal planes. For the specimens where XZ and YZ were approximated by cut A, B and C, it was decided to measure the strain ratio in XY to test the compatibility of the ratios on sections A' and B'. A major factor in the failure of the application of the Roberts and Siddans method was the problem of sections at about 45° to bedding and the lack of information from similarly oriented sections in the undeformed state. For future routine studies using the Elliott method it is essential to use principal sections only and to discard sections where X cannot be determined accurately. Also more data is needed from undeformed rocks to test the variability of the initial fabrics in several orientations.

The XY thin sections for specimens 44367/72/73 were cut, measured and the results presented in the form of Elliott plots (Figure 7.29). Original boundaries of detrital grains were very poorly delineated in the XY section of specimen 44368 which was, therefore, not further considered. The three principal sections for the three remaining specimens were studied using the methods outlined previously. Several strain values were possible for some sections, but, the additional sorting criterion provided by the internal check, allowed firm decisions to be made in each case.

TABLE 7.8

	XY	YZ	XZ
44372	$^\wedge S_0 \approx 80^\circ$ Moderate reversed shape fit, good S_0 fit Viking $\perp$ S_0 at $\varepsilon = 0.126$ Penrith S_0 Good shape + S_0 fit at $\varepsilon = 0.089$ $\varepsilon_s = 0.089$ $R_s^s = 1.19$	$^\wedge S_0 = 10^\circ$ Very good fit to Viking D $\parallel$ to S_0 $\varepsilon = 0.210$ $\varepsilon_s = 0.210$ $R_s^s = 1.52$ (1.69)	$^\wedge S_0 \approx 80^\circ$ Good shape fit and excellent S_0 to Viking $\perp$ S_0 at $\varepsilon = 0.310$ $\varepsilon_s = 0.310$ $R_s^s = 1.86$ (1.38)
44367	90° to S_0 All fits within the range of $\varepsilon = 0.020$ to 0.043 to have a good to moderate S_0 fit. $\varepsilon_s = 0.031$ $R_s^s = 1.06$	90° to S_0 Moderate shape fit, very poor S_0 with Viking $\perp$ S_0 at $\varepsilon = 0.105$. Moderate shape and very good S_0 fit to Viking $\perp$ S_0 at $\varepsilon = 0.167$. $\varepsilon_s = 0.167$ $R_s^s = 1.40$ (1.42)	$\parallel$ S_0 Moderate shape fit to Viking D at 0.180. Similar values for other bedding $\parallel$ fabrics. $\varepsilon_s = 0.180$ $R_s^s = 1.43$ (1.45)
44373	$^\wedge S_0 \approx 80^\circ$ Good S_0 trace and shape fit to Viking $\perp$ S_0 at $\varepsilon = 0.055$ $\varepsilon_s = 0.055$ $R_s^s = 1.17$	$^\wedge S_0 \approx 80^\circ$ Good shape and moderate S_0 fit to Viking $\perp$ S_0 at $\varepsilon = 0.080$ $\varepsilon_s = 0.080$ $R_s^s = 1.17$ (1.06)	Low angle to S_0 Good shape fit to Viking A $\parallel$ to S_0 at $\varepsilon = 0.140$ $\varepsilon_s = 0.140$ $R_s^s = 1.32$ (1.22)

Values in brackets after R_s refer to values first estimated in the Roberts and Siddans method.

Table showing the shape factor fits used on the principal planes in combination with the internal check X/Y . Y/Z = X/Z.

The procedure is outlined in Table 7.8 where the chosen strain ratios are compared with those previously determined by the Roberts and Siddans method. There is good agreement between these values, further emphasising that the problems were mainly caused by the C' sections which were invariably at an intermediate angle to bedding. It is fortunate that in the specimens measured most of the principal planes were either at a high or low angle to bedding. Obviously this relationship will not be found universally and more complete initial fabric information must be gathered.

The three-dimensional strain states for all the quartz arenite specimens that were completely analysed are shown on a log/log deformation plot in Figure 7.30. All except one fall in the flattening field and there are no great disparities in the type of strain pattern displayed. The other strain result lies on the plane strain line.

7.9: Summary and Conclusions on Strain Analysis Methods

This study has emphasised the need to utilise all three principal sections when measuring tectonic strain, whichever of the techniques is employed. The Elliott method was chosen as it appeared capable of the most general application. The choice was reinforced during the analysis of unstrained sedimentary fabrics as the information gathered here showed that the basic assumptions of the other techniques were inadequately based. It appeared that results from the competing methods were quite inaccurate and until satisfactory tests can be made they were passed over. Taking note of the initial fabric data and its application to the deformed oosparites, the published Elliott method was modified.

Essentially, fabric patterns from similar sedimentary environments were compared with the deformed examples in order to locate the initial circle points. It was determined that the following factors had to be allowed for in the strain assessment:

1) the angular relation of bedding to the section of the deformed and undeformed material,
2) the location of the ICP as close as possible to the line of maximum elongation of the deformed section,
3) the goodness of shape fit between the deformed and undeformed plots,
4) the minimisation of the angular difference between the S_0 traces of the undeformed plot and the restored deformed section.

A successful application of the modified approach was made on the deformed quartz arenite of the Southern Franklands and, providing all the above parameters are considered, the method can be generally applied where appropriate initial fabric data is available.

Further work must involve the more extensive study of pre-deformation fabrics. Another outstanding aspect is the need to test validity of the alternative methods. Data of the kind gathered in this project can be applied to this latter task.

CHAPTER EIGHT

SUMMARY AND CONCLUSIONS

The final chapter intends to be an integration and resume of previously given facts and interpretations. The discussion will be structured essentially around a geological history. Such a chronological approach, however, is unsuitable for some aspects of the work and these will be dealt with somewhat separately.

8.1: Sedimentation

The great bulk of the pre-tectonic rock types were sedimentary. Metamorphosed quartz arenite and mudrock dominate with only a very minor carbonate representation. Sedimentary structures within the quartz arenite indicate deposition on a tidal shelf though textures are more akin to an aeolian environment. Considering the prevailing conditions on land in the late Precambrian wind transport is expected to have been more important than at present. It appears that the sediment eventually incorporated into the shallow-shelf sea deposits had a long prior history of desert influences. Meta-mudrocks rarely preserve sufficient numbers of sedimentary structures to allow a sound environmental interpretation. Local areas, however, demonstrate features which may either represent a tidal-flat or a delta-front platform. The former hypothesis is favoured by the regional interpretation but more critical information is required to confirm the interpretation; the strain and metamorphic state of the rocks may prevent the recognition of such features.

Considerable care was taken in the identification of true sedimentary structures which if mistakenly considered to be tectonic can lead to radical differences in regional and local interpretations.

Folding has been demonstrated during penecontemporaneous deformation of cross-bedding. Such folds are morphologically identical to tectonic examples and the problem of recognition may be compounded in areas where layering and fabrics are parallel. Soft-sedimentary deformed cross-bedding may lead to asymmetric first-generation tectonic folds being placed in a later deformation phase if they occur on the common limb of the F_1. A consideration of untectonised equivalents demonstrates some of the possible problems and makes the worker in a deformed terrain aware of complications. A further example of the need to study undisturbed sedimentary environments is the varying interpretations placed on clastic dykes sub-parallel to cleavage. The majority of such cases appear to be best explained in terms of strain superimposed on typical examples of clastic dykes seen in uncleaved rocks.

8.2: Nature of the D_1 Event

When comparable quartzites are examined, D_1 recrystallisation is more advanced in the northern zone of the Franklands and Wilmots than at the south eastern end. The strain quantification carried out in quartz arenite showed a direct relationship between the degree of recrystallisation and strain which, therefore, enabled qualitative assessment of D_1 strain to be based on fabric. A general decrease in strain is indicated from north to south and, though sedimentary structures are found throughout, evidence of sedimentary textures progressively becomes clearer towards the south where an average of 20% to 25% shortening in Z is estimated. In the north, S_1 is almost everywhere parallel to bedding yet south of Cleft Peak S_1 is commonly at an angle to bedding. Major D_1 folds are best displayed within the low strain zone; they are fold nappes as it appears that their lower limbs are fairly extensive. The maximum limb length involving inversion of stratigraphy has been proved to be of the order of 5 kilometres.

There is, in D_1 a constant pattern of east facing folds (Figure 8.1 and 8.2). All large-scale D_1 folds, therefore, indicate that they were produced by overriding from the west or southwest and their asymmetry when viewed northwards along the mountain chain must be Z type. Despite this information it is impossible to give the orientation of enveloping surface to the largest-scale, first-generation folds. Thrusting is associated with the folding and the only clearly defined thrust is considered to be synchronous with fabric development.

In the eastern Frankland Range both tight and isoclinal major D_1 folds have strain distributions expected from the even distribution of concentric shearing strain through the layering in the limbs. An overall, homogeneous flattening of approximately 25% is also indicated from studies of modified cross-bedding. Recumbent fold nappes are known in only slightly strained rocks (for example, the Helvetid nappes of the European Alps, see Hobbs, et al., 1976, p.415) and several authors (Ayrton and Ramsay, 1974; Wood, 1973) have shown that much of the strain developed in these structures is later than the nappe formation events though Tan (1976) has presented evidence against this sequence of events for one region. Many of the Prealpine nappes appear to have undergone bulk translation probably in slices bounded by discrete thrust surfaces. Lower levels of the nappe pile are flattened under the weight of overriding high level nappes and there may be a considerable time gap between emplacement and strain (Ayrton and Ramsay, op. cit.). A sequence similar to that outlined above is tentatively suggested by observations made on the stretching directions (X) developed in D_1.

Below the thrust in the eastern Frankland Range, the X direction is usually about 30° away from the major fold axes. Above the thrust, in perfectly isoclined D_1 folds, X is usually at a higher angle to the fold axes and may be 90° away. If the deformation occurred during tectonic transport, a more consistent X orientation may be expected in these major folds with a high angle to the axes being preferred. The results favour flattening after translation with the variation in the X orientation reflecting local constraints. The northern portion of the Frankland and Wilmot Ranges, which is characterised by S_1 parallel to layering, may represent deeper tectonic levels which were flattened under the load of the nappes above. Fold mechanisms have been mainly investigated in the major tight chevron folds below the inferred thrust of the eastern Franklands. Above the thrust and associated strain concentration zone, the rounded isoclines again show low strain states. Presumably the rounded nature of the closure allowed tangential longitudinal strain adjustments to be uniformly spread at low values through this part of the fold. The observed large isoclines have nearly constant curvature with perhaps a slight increase at the hinge.

In the immediate post-D_1 situation the major folds are considered to have been gently inclined or recumbent. Small fold nappes with limb inversions of about 5 kilometres have often been considered to develop as near symmetrical, steeply inclined or upright folds which then collapse under their own weight (Roeder, 1977; Roberts, 1974). If the Frankland's folds follow this pattern then their pre D_2 attitude may have been somewhat variable but no root zone is in evidence where steep dips have dominated throughout the deformational history. Gravity is considered to be the principal force responsible for the emplacement of the D_1 folds. The very low strain states argue against gravity flow and strongly favours gravitational gliding probably in units bounded by fairly discrete displacement discontinuites. The nature of the strain symmetry in the thrust described here has not been characterised. It probably developed as essentially a plane strain type during simple shear and was subsequently modified by a flattening strain. Close to the thrust in the highest strain area measurable, a total flattening pattern was recorded. Basement is not seen and its role during orogeneses is impossible to evaluate. Other aspects of the tectonics of the first deformation are also uncertain. The consistent facing of D_1 folds and their inferred asymmetry may have been caused by sub-horizontal translation of an undistorted suprastructure (Fyson, 1971). From the characteristics of the eastern Franklands this area may be expected to be on the lower boundary of the suprastructure. The position of this interface is largely controlled by the onset of metamorphic reactions which release volatiles and thereby increase creep rates (Holland and Lambert, 1969). Textural evidence indicates low greenschist facies conditions during D_1 and the disparity of behaviour between detrital quartz grains and secondary overgrowths supports the operation of temperatures low enough to prevent dehydration.

In discussing the first generation of folds to be recognised in a preliminary study of the region, Powell (1969a) states, "The bulbous concentric style of the folds and the absence of well-developed cleavage indicate that this movement may have occurred early in the lithification history of the sediments". Also during the present study, clastic dykes were found to be very close to cleavage where S_1 was at an angle to S_o. It could, therefore, have been argued that, at the time of D_1 folding, the rocks of the Frankland and Wilmot Ranges were not completely lithified and that tectonic dewatering was responsible for the production of the first fabric. Structural analysis, however, showed that the majority of the first generation folds recognised by Powell (1969a) in fact folded a tectonic fabric, had a generated axial-plane crenulation cleavage and folded layering which had been boudinaged. The style of his 'first folds' was, therefore, considerably influenced by previous tectonism and may not be used to infer the state of lithification at the time of the first deformation. D_1 strain has caused recrystallisation of secondary overgrowths in quartz arenite which again demonstrates the lithified nature of the sedimentary pile at the time of deformation.

A detailed analysis of the clastic dykes demonstrated that, though they had in outcrop a strongly planar form and a sub-parallel relationship to cleavage, they showed strongly modified compaction folds transected by S_1. Original sill-like intrusions were found as strongly contorted interconnecting branches. These features were best accounted for by considering the effects of reasonable amounts of strain on clastic dyke patterns commonly described from untectonised terrains. This view was strengthened by the presence of clastic dykes where S_o paralleled S_1; a situation that cannot be generated by tectonic dewatering. During the last five years observations on clastic dykes in deformed terrains have become quantitative. In the type area for tectonic dewatering it has been shown that the dykes are not parallel to cleavage and yet it was the belief that dykes and cleavage were parallel that germinated the hypothesis. It is interesting to speculate on whether the hypothesis would have been formulated if the correct observations had been made initially.

Recent arguments from proponents of the tectonic dewatering hypothesis now allow some divergence of dykes and cleavage. Powell (1976) considers that statistical parallelism of dykes and cleavage is sufficient evidence to prove the hypothesis. This, however, is exactly the expected situation if planar, bedding-perpendicular dykes were folded and cleaved. If the folds were approximately symmetric and dykes evenly distributed then during strain the dykes would converge on the cleavage and have a symmetric distribution about cleavage. It is concluded that there is no evidence to unequivocally support the tectonic dewatering hypothesis. Because in many regions it has been shown to be impossible for this mechanism to account for cleavage (e.g. Wood, 1974; Holeywell and Tullis, 1975; Beutner, et al., 1977) it is considered that it may only be responsible for the actual initiation of cleavage in a few restricted environments (see Powell, 1976).

8.3: The Metamorphic Climax

Textural evidence indicates that the peak of metamorphic crystallisation occurred immediately before D_2 and near peak temperatures must have obtained during D_2. Small garnets appear to be mainly pre-D_2 whilst larger garnets show pre-and syn-D_2 relationships. Some porphyroblasts grew through to the end of D_2 and perhaps a little after. In the case of pre and syn-tectonic crystallisation, it is usually the pre-tectonic stage that predominates.

Pressure and temperature conditions in the highest grade zone at the peak of metamorphism have been estimated by combining several lines of evidence. Amphibole and phengite chemistry taken in conjunction with mineral assemblages viz. the association of pure albite and almandine garnet, the presence of chloritoid, and the formation of garnet before biotite, indicate temperatures between 400^{o} to $500^{o}C$ at a load pressure of about 6 kb. On a regional scale within the Tyennan Geanticline there are no syn - orimmediately post-orogenic granites and other metamorphic studies have indicated a high pressure with low to intermediate temperature regime. General indications are for a geothermal gradient between those of the Barrovian and Sanbagawan regions.

Except for those of the smallest scale, D_2 folds show a consistent sense of asymmetry throughout the region (Figures 8.1: and 8.3); F_1 and F_2 are also coaxial. This pattern can be used to infer a direction of over-thrusting which parallels that for the major transport direction of the first folds. Such a "symmetry continuation" is often taken to mean that the first two deformation phases are the response to a single continuous episode of over-riding from one direction. It is also possible that the tectonism took place in pulses separated by periods of inactivity. This latter suggestion is favoured by several features of the D_1 to D_2 interval in this region and by comparisons with other studies of comparable situations. Doleritic dykes, tholelitic in composition, are considered to have been intruded before and possibly during the early part of the metamorphic climax. Whilst intrusion is possible in a regional compressive stress field, the timing indicates some break between D_1 and D_2. Some time lapse is also supported by the overprinting of S_1 by porphyroblast growth which is mainly pre-S_2. The step-wise process of folding, translation and then strain during D_1 shows that the first portion of the D_1 to D_2 sequence is not continuous. In the Eastern Alps, the major period of overthrusting occurred some 25-30 m.y. before the peak of metamorphism (Oxburgh and Turcotte, 1974) demonstrating the existence of large time gaps in some regions between episodes of nappe formation and high grades of metamorphism. The closer these two events are in time, the greater is the requirement for heat input from the mantle (Richardson and Powell, 1976) andthis may be excessive if D_1 and D_2 are virtually continuous. Chemical variation of phengite grains from S_1 to S_2 is gradual which reflects continuous metamorphic changes. Sudden changes in chemistry in this interval are lacking and may have been expected if temperatures had increased rapidly in a narrow time interval. The balance of evidence favours an unspecified hiatus between D_1 and D_2 though, considering the usual climactic nature of orthotectonic orogeny, the break is unlikely to be considerable. D_2 folds are the result of overriding from the west and southwest and probably have same fundamental cause as the D_1 fold nappes. A source zone and a region of uplift is indicated to the west and southwest.

From Terminal Peak to Coronation Peak there is a consistent plunge to the major folds which exposes some five kilometres in depth of profile. Considering the plunge variations from Coronation Peak to the Gordon Dam, another one kilometre of depth may be exposed. Because the higher grade zones are exposed in the north, it seems unlikely that this regional plunge developed after the metamorphic peak. Direct evidence for the production of the plunge is, however, lacking.

8.4: Structural Evolution Post The Metamorphic Climax

After D_2 there is no evidence of porphyroblast growth though quartz locally shows marked modification through strain and recrystallisation. The fabrics S_3, S_4, and S_5 clearly reflect the operation of lower P/T conditions. Chemical changes, related to the formation of S_4 overprinting S_2, are sharp and indicate a considerable fall in temperature from D_2 to D_4. On the basis of the present evidence it is impossible to say whether temperature fell gradually from D_2 to D_4 or if there was a trough in the temperature curve with a small rise to D_4. During a gradual decrease of temperature, mineral composition adjustments may be sluggish and reaction rates may only increase to produce noticeable changes when the strain energy of a deformation pulse is sufficient to exceed the activation energy barrier of the rate determining step.

Locally, D_3 produced folds with amplitudes close to one kilometre but they are generally on a scale of less than ten metres. The larger folds occur in flaggy quartzite/micaceous quartzite alternations and D_3 effects are very minor in pure quartzite zones. D_3 folds tend to show a consistent sense of asymmetry except for reversals on the common limbs of the highest order folds. Their geometry indicates subhorizontal overriding in an opposite sense to the D_1 and D_2 folds. The D_3 folds are nearly coaxial with earlier generation folds. Because of the intermediate size of both D_2 and D_3 folds and the likely near recumbent nature of the first folds pre-D_2, it seems that F_2 and F_3 were superimposed on approximately flat-lying layering. Axial surfaces to both fold generations were probably inclined at less than 45^o to the horizontal (Figure 8.3c) and in incompetent rocks S_2 and S_3 were shallowly dipping. The conjugate arrangement of F_2 and F_3 axial surfaces may suggest synchronous development but overprinting is always consistent and the nature of the microfabrics is quite different.

D_4 may be described as the paratectonic stage of the evolution of this deformed terrain. It is characterised by major, upright or slightly inclined folds which have little or no plunge. The largest proved D_4 fold has an amplitude close to four kilometres and this generation is the major influence on the attitude of layering. Large-scale, near-coaxial rotation of all pre-existing structures took place in D_4 (Figure 8.3) and the effects of this deformation phase are ubiquitous. Strain in D_4 does appear to markedly increase towards major fold closures and may be low to moderate on the limbs. Wherever S_4 is intensely developed there is an increase in the development of S_5. This latter surface is a weak crenulation usually associated with folds whose amplitudes are less than one metre. F_5 axial surfaces are inclined, on average, at 45^o and folds of this generation may be genetically related to D_4. Where strong vertical stretching has taken place D_5 structures are more common and they may reflect vertical shortening of the column which extended beyond its capacity to support itself.

After the minor structures of the D_5 event, a series of small-scale conjugate kinks were imprinted on the rock fabric. Perhaps the most obvious major structure of the whole area, the rotation of D_1 to D_5 and possibly two conjugate kink generations, occurred quite late in the structural evolution of the region. D_1 to D_5 hinge lines and planar fabrics show north/south trends at the Knob/ Serpentine and east/west trends at Terminal Peak. A major rotation about a vertically plunging axis is indicated. Corbett (1970) described the Lake Edgar Fault as a major north/south trending dislocation which lies about five kilometres to the east of the Scotts Peak Dam (Figure 1.2). From reconnaissence regional mapping, Corbett (op. cit.) proposed that there had been considerable pre-Ordovician transcurrent movement along the Lake Edgar Fault and that the swing of the Frankland and Wilmot Ranges was related to this event. The above interpretation indicates that D_1 - D_5 and two conjugate kink phases are pre-Ordovician. The work of S. J. Williams (1976), and Maclean and Bowen (1971) shows that some minor kink folding within the basement may be attributed to the widespread Middle Devonian Tabberabberan Orogeny. Little information can be gained on the timing of the discontinuous deformation described from the region of the Gordon Dam except that it is later than the cleavage forming events.

8.5: Microfabric Summary

Dislocation deformation mechanisms operated during D_1 in quartz arenite almost to the exclusion of other means of achieving large strains. The amount of recrystallisation can be related to the strain indicating the importance of dynamic recovery and recrystallisation. Temperatures during D_1 were low enough to prevent dehydration of quartz overgrowth on detrital grains and, under similar strain conditions, very pronounced differences in recrystallisation behaviour are related to variations in hydrogen content. Strong hydrolytic weakening occurred in the overgrowth. The first fabric in micaceous lithologies is a perfect dimensional preferred orientation of micas with quartz grains displaying various degrees of elongation. Little evidence is preserved to indicate the mechanisms which operated to produce this fabric.

In post-D_1, pre-D_5 events grain-boundary diffusive mass-transfer processes become important. The impetus for this mechanism was often stress variations resulting from microfolding but significant differentiation occurred without folding to form a pure 'pressure solution cleavage'. New grains produced in D_1 often were recycled through a deformation subgrain/new grain sequence in D_2 to D_4.

Crystal plasticity allowed the change of shape, followed by dynamic recrystallisation. S_2 in micaceous rocks is often a well differentiated crenulation cleavage; it formed near peak temperatures which must have facilitated differentiation with a greater degree of microfolding.

Microfolding morphology, particulaly in D_4, is closely related to rock type. In mica-poor quartzite ($<5\%$ mica), deformation produces folds whose wave length depends upon mica content.

In rocks of low mica content crenulations are coarsely developed microfolds with a 2 cm spacing. However, the structures show a gradation in their spacing according to mica content and in phyllite microfolds have a 1 mm spacing or less.

8.6: Fabric Strain

Elliott (1970) showed that on theorectical grounds a wide variety of initial sedimentary fabrics were possible when dealing with strain markers such as ooids, oncoids, accretionary lapilli, quartz grains, and pebbles. The complexity of the initial state called into question the validity of using computerised strain analysis methods which rested on rather limited assumptions about pre-tectonic fabrics (Dunnet and Siddans, 1971; Matthews, Bond and Van den Berg, 1974). The tectonic shape factors of the Elliott method appeared capable of giving a large amount of information concerning the initial fabric type and hence the strain analysis was not based on unsubstantiated assumptions. Unfortunately the theoretically derived sedimentary shape factors were not easy to recognise in practice and further investigation indicated that the different patterns may not be very distinct.

The initial aim of the project was to quantify strain in quartz arenite from the Frankland Range but it eventually turned towards an investigation of the strain analysis methods themselves. Clearly a major problem in this work was a lack of characterisation of the nature of initial fabrics in a form appropriate to the structural method. Direct measurements of potential strain markers was carried out on specimens from terrains where the effects of strain were negligible. The measurements linked axial ratio and long axis orientation for individual strain markers; the key parameters whose combined behaviour during strain is of most interest. Many fabrics, though being approximations to the unimodal pattern, were shown to be imbricate. This must cause errors if the method of Matthews et al. (1974) is used which assumes bedding symmetric fabrics. The same assumption is used in the most commonly applied option of the Dunnet and Siddans (1971) method viz. the minimum difference criterion. Measurements of unstrained rocks also showed that inspection of shape factors could not confidently identify the initial circle point from the characteristics of plot shape alone (the method of Elliott).

The investigation demonstrated that it was vitally necessary to study the unstrained equivalents of any terrain or to have available suitable information from such areas. Pre-tectonic textures and structures greatly influence behaviour of materials during tectonism and the effect of a sedimentary fabric may be noticeable until high strains. These conclusions apply equally well to thin section scale grain 'shape factor' studies or outcrop scale investigations of modified pre-existing features.

The present study was carried out on quartz arenite specimens where detrital quartz grain boundaries were clearly outlined by dust trails. Apart from any possible minor pressure solution strain in the matrix, deformation was achieved by crystal - plastic flow. Tectonic strains were recorded by evaluating the contribution to the final state of the sedimentary fabric. Throughout it was assumed that the cleavage plane was the XY plane of the finite strain ellisoid. The present study cannot add to the present discussion on this matter (see Borradaile, 1977) except to say that the information gathered here is not incompatible with the proposal. Strain analysis using complex initial fabrics cannot be used to prove that cleavage is parallel to XY. Another, as yet little investigated, assumption is that the initial markers are sufficiently close to elliptical in shape so as not to invalidate the transformation equations.

A methodology evolved through several stages essentially based on the elegant plotting method of Elliott. In the first instance the vestiges of the sedimentary features in a terrain are analysed to determine as much as possible about the environment of deposition. Sedimentary shape factors are characterised by measurement of similar materials from unstrained regions. The shape factors are compared with those measured on principal planes in the deformed rocks, where possible matching sections with similar angular relations to bedding. During the process of pattern fitting the origin of the unstrained specimen is kept as close as possible to the inferred maximum elongation of the deformed specimen and, in sections at an angle to bedding, the position of the bedding trace is used to account for imbrication. Results for the strain ratio in a single plane must be analysed in the light of those obtained for the other two principal planes. It is considered unsound to rely on only two results, with no means of independently checking internal consistency, when initial fabrics may be so variable.

Several recent publications have used variations on the above theme. Tobisch et al. (1977) restore deformed fabrics to the mean of determinations found in unstrained rocks using appropriate section/bedding dihedral angles. It may be possible to improve this type of approach as volume concentration of some compaction sensitive markers (e.g. accretionary lapilli) appears to be related to the compaction strain. If this type of information is considered a better restoration is feasible. Huddleston (1976), however, uses the Elliott method butsimply assumes the fabrics were initially random and takes the centre of the plot on each plane as the initial circle point. Further details of this study have yet to be published though it is claimed that this assumption leads to the simplest relationship between strain and the tectonic fabric.

Seven quartz arenite specimens were completely analysed during the strain measurement program for the southeastern Franklands. All results except one belonged to the flattening field with average extensions of +10% in the Y direction. Shortenings in the Z direction varied from 18% to 36%.

8.7: Structural Correlation in Southwest Tasmania

Regional correlations of fold phases within the metamorphosed Precambrian of southwest Tasmania have implications for the geological history of the Franklands and Wilmot area. A predominant feature of the region is a post-crystallisation upright folding event (Maclean and Bowen, 1971; Williams, S.J., 1976; Williams, P.R. and Corbett, E.B., 1977) which is likely to be approximately synchronous. At Port Davey a low-grade, pre-Ordovician, metasedimentary sequence overlies the metamorphic sequence; the first regional cleavage developed in the upper sequence passes directly into a crenulation in the basement (Williams and Corbett, 1977). The basal conglomerate contains blocks of the metamorphosed Precambrian and a considerable time break can be inferred between D_1/D_2 with the associated metamorphic peak and the later structural events. The metamorphosed Precambrian insouthwest Tasmania is, therefore, the result of two separate, pre-Ordovician, tectono-metamorphic cycles each of which were polyphase (c.f. Tobisch et al., 1970). From the description of the Port Davey area (Williams and Corbett, 1977) it is uncertain whether or not it is the widespread upright fold event of the basement which is synchronous with the first regional cleavage of the overlying sequence; deposition of the latter must be at least post - D_2.

Other mapped regions in southwest Tasmania show several marked similarities in structural characteristics (see Table 8.1). The first deformation has generally given rise to well-developed tectonic fabrics almost everywhere parallel to compositional layering. The peak of metamorphism is generally placed between D_1 and D_2 in regions where detailed structural and textural analysis has allowed clear separation of deformation phases. Folds of the second generation are tight to isoclinal in sequences of quartzite/ phyllite interlayering and are associated with pronounced fabrics. D_2 appears to have formed very close to the peak of metamorphism. D_1, D_2 and D_4 of this study have the most important effect on fabrics and have been correlated with the main regional events D_1, D_2 and D_3 events of Williams and Corbett (1977) who note the presence of an extra phase between the main regional events. Maclean and Bowen's (1971) second phase folds at the Davey River are "tight to isoclinal with vertical axial surfaces:. Their characteristics are very similar to F_4 of the thesis region and this correlation is further strengthened by the nature of D_1 at the Davey River which is described as possibly containing two textural events prior to D_2. It seems evident that Maclean and Bowen were not

THIS WORK	PORT DAVEY Williams, P.R. and Corbett, E.B. 1977	DAVEY RIVER Maclean and Bowen, 1971	SENTINELS Williams, S.J., 1976	FRENCHMANS CAP Spry, 1963b	GORDON RIVER DAMSTIE Powell, 1969a
D_1	D_1	F_1	D_1	F_1	
First phase of deformation gives rise to widespread tectonic fabrics parallel to compositional layering.					
Metamorphic peak is immediately prior to	Two growth periods of garnet pre-D_2	F_1 is described as a complex event and almost certainly includes D_1 and D_2 of the Franklands.	Static metamorphism		G_1 This appears to be mainly equivalent to D_2 of the present study.
D_2	D_2		D_2	F_2	
	A local event is recorded at Port Davey between the main regional events.	Metamorphism is between F_1 and F_2			
D_3	?Equivalent to D_3 of the Franklands? UPLIFT EROSION OF METAMORPHIC BASEMENT AND DEPOSITION OF SEDIMENTS BETWEEN D_2 and D_3.			A composite F_3 of several events perhaps including all post D_2 phases of other workers and some local events.	
D_4	D_3	F_2	D_3		G_2
Major upright folds	Upright folds	Major upright folds	Major upright folds		
Later minor events and large scale 'drag' on trancurrent fault.	Later minor kink-bands and chevron folds	Later minor events.	Minor D_4 and two Tabberabberan kink sets.		G_3

TABLE 8.1 CORRELATION OF STRUCURAL EVENTS IN SOUTHWEST TASMANIA.

able to distinguish the effects of D_2 (and possibly D_3) as determined for the Frankland Range. The first structural analysis carried out on the thesis area (Powell, 1969a) reported only one phase of deformation before the main upright event. From the description of Powell's F_1 it appears that most fold examples measured by him belong to the present D_2 category and that their post D_1 nature was not recognised. In an earlier report (Boulter, 1974a) the effects of D_3 were only weakly developed in the regions studied to that time and they were not distinguished as a separate event. The D_3 of Boulter (1974a) becomes D_4 with the addition of the intervening phase.

8.8: Observations on Some General Aspects of Tasmanian Precambrian Geology

Approximately one fifth of the State of Tasmania is occupied by Precambrian rocks. The bulk is found in two separate large regions of outcrop; the Rocky Cape Block and the Tyennan Geanticline (Figure 8.4).

Both areas have contrasting tectonic styles with the former displaying a predominance of open to tight almost upright folds and low grade metamorphism whilst the latter has large areas with strong fabrics parallel to layering and greenschist to lower amphibolite fabrics metamorphism. Correlates of the Tyennan Geanticline rocks are found in thrust contact with material of the Rocky Cape Block at Goat Island (427446, Figure 8.4) but no other contacts are known. Despite the lack of clear relationships between the two units it is commonly believed that the bulk of deformation and metamorphism of the Tyennan Geanticline (Frenchman Orogeny) predates the deformation in the Rocky Cape Block (Penguin Orogeny) (e.g. Williams, E., 1976). Little work has been carried out to determine the age of these metamorphic events by radiometric means though available evidence points to a close temporal relationship.

Within the Rocky Cape Block, dolerite dykes intruded the sequence during the first folding episode (Gee, 1967) and these have been dated by K-Ar whole rock methods at 720 m.y. (Richards, J.R. in Solomon and Griffiths, 1974, p.21). D_1 in this unit, therefore, has a minimum age of 720 m.y., a date which presumable reflects the cessation of argon diffusion. The age may also have been modified during the Tabberabberan Orogeny which is known to have caused near total readjustment of many K-Ar mineral systems within the Precambrian rocks (McDougall and Leggo, 1965, p.319). Further evidence on the age of the Penguin Orogeny is to be found on King Island about 90 kilometres north-northwest of mainland Tasmania. Here granitoid rocks have intruded metasediments after the first phase of regional folding (Cox, 1963). The meta-sediments can be traced almost continuously from the Rocky Cape Block on the mainland and it appears that the granitoids have intruded the sequence after regional D_1. McDougall and Leggo (1965) carried out Rb/Sr whole rock and mineral determinations on the King Island granitoids but at the time of this work the geology was poorly known and they appear to have sampled two plutons to produce and isochron. The above authors (p.319) 'conclude that emplacement most probably occurred about 750 m.y. ago. However, the possibility that the age is approximately 835 m.y., as suggested by the whole rock data, must be borne in mind'. Recent Rb/Sr studies, both whole rock and mineral, have indicated that the peak of metamorphism in the Tyennan Geanticline (the Frenchman Orogeny) occurred at about 800 m.y. ago and that a superimposed event took place in the 550 to 630 m.y. range (Raheim and Compston, in prep.). This latter event cannot be equated with the Penguin Orogeny of the Northwest Coast and many features point to a similar time of formation for the Frenchman and Penguin orogenies. D_1 in both areas need not be synchronous but the episode of dolerite intrusion (syn-D_1, Rocky Cape Block; post-D_1, pre-metamorphic climax and D_2, Frankland/Wilmot Ranges) may provide a reference plane. First generation folds in the Tyennan Geanticline show clear evidence for derivation from the west and first folds in the Rocky Cape Block have axial surfaces inclined to the west. Certainly tectonic transport on a regional scale is from the same direction and may be related to the one period of crustal disturbance. It is tempting to suggest that the rocks of the Rocky Cape Block are the supracrustal equivalents of the Tyennan Geanticline rocks which formed an infrastructural regime in the manner described by Fyson (1971).

It is interesting to note that little evidence of strong deformation at about the time of the Frenchman Orogeny is to be found on the mainland of Australia. In a Gondwanaland reconstruction for the southwest Pacific (Griffiths, 1974) it can be seen that several late Precambrian sequences in Northern Victoria Land form a near continuous zone with the Tasmanian Precambrian. The Wilson group in this nearly contiguous zone has a reported metamorphic age of 770 $\pm$ 20 m.y. (Rb/Sr determination) and grades range from greenschist to amphibolite facies (Nathan and Skinner, 1970). This group is a very strong candidate for correlation with the rocks and main deformation/metamorphism in the Tyennan Geanticline.

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APPENDIX

ROCK SPECIMEN COLLECTION

The following list includes specimens that were collected specifically for the present thesis work. Other material has been taken from pre-existing collections and this is noted in the text.

For sample locality grid coordinates, unless otherwise stated, the Grid Zone Designation is 55G and the 100,000 metre square identifying letters are DN.

39166 Grid ref. 16706740; preparations, T, PT, CA, PE; quartz - phengite- chlorite - calcite schist, Precambrian from Gordon Power Station.

The group of apecimens with numbers between 40585 to 40712 were collected during a study of clastic dyke/cleavage relationships in the Precambrian rocks of the Frankland and Wilmot Ranges, Southwest Tasmania. Not all have been cited in the thesis. They were collected from remote outcrops or from roadside localities. The latter suite was accumulated before the section was used for student excursions and gives a representative collection. These specimens have been referred to in the thesis and in Boulter (1974b).

Locality (1) of Figure 2.8 is the outcrop at the 47.5 mile post on the Gordon River Road between grid coordinates 19126423 and 19716442. Locality (2) of the same Figure is the saddle between Cleft Peak and Greycap were specimens were collected along a narrow ridge between 25964194 and 26234191. Locality (3) is 1.8 kilometres down the Serpentine Road at 16796503.

The following specimens were all taken from the centre of the saddle between Cleft Peak and Greycap, Frankland Range, S.W. Tasmania.

40585 Planar clastic dykes perpendicular to S_o and parallel to S_1. Section cut parallel to S_o to show S_o/ dyke intersection pattern.

40586 Planar clastic dykes perpendicular to S_o and parallel to S_1 with contorted branching intrusives.

40587 as 40586

40588 as 40586

40589 as 40586, dyke margins are extremely irregular.

40590 as 40586

40591 as 40586, low angle cross cutting cleavage (S_4).

40592 as 40586, T

40593 Planar clastic dykes perpendicular to S_o and parallel to S_1 and compositional layers.

40594 as 40586, irregular dyke margins.

40595 as 40594

40596 as 40594

40597 as 40594

40598 as 40586, dyke refraction and tendency for transecting cleavage (S_1) to follow.

40599 as 40598

40600 Penetrative S_1 in silty unit with planar clastic dyke parallel to cleavage.

40601 Core of an F_1 closure with multiple, tabular clastic dykes sub-perpendicular to S_o following S_1.

The following specimens are from the S.E. end of the pelitic outcrop in the saddle between Cleft Peak and Greycap, Frankland Range.

40602 Short planar clastic dykes parallel to a penetrative cleavage (S_1) cutting compositional layers at 30^o.

40603 as 40602

40604 as 40602

40605 as 40602, direction of intrusion is indicated by bowing of internal laminations.

40606 as 40602

40607	as 40602
40608	as 40602
40609	Short, multiple planar dykes parallel to a penetrative cleavage. Compositional layering of mud/silt showing partially preserved sedimentary structures.
40610	as 40609, T
40611	as 40609, T
40612	as 40609, T
40613	as 40609
40614	as 40609
40615	as 40609
40616	as 40609
40617	as 40609

The following are from the centre of the saddle between Cleft Peak and Greycap, Frankland Range. (Except 40621)

40618	Dominantly planar habit of the clastic dykes // to S_1. Margins less irregular than many examples but interconnecting branches almost, large T ptygmatic.
40619	Irregular clastic dykes sub-parallel to S_1. Strongly refracted in part, large T.
40620	Dominantly planar clastic dykes with irregular margins and strongly contorted branches.
40621	Quartzite clastic dykes in pelite from the Knob Quarry, Gordon River Dam site, Strathgordon.
40622	as 40620
40623	Thick planar dykes parallel to S_1 with intra-folial folds where S_1 is axial planar. T

The following are from the Gordon Road section 47.5 miles from Maydena.

40624	Multiple, contorted, tabular clastic dykes. Those at high angle to S_o show ptygmatic folding. Low angle intrusives are generally planar. Structureless feeder zones indicate liquefaction origin.
40625	as 40624

40626 as 40624, large T

40627 as 40624

40628 Thick clastic dykes with pronounced structureless feeder zones, large T

40629 Multiple, contorted clastic dykes where S_1 is parallel to S_o and the effects of S_3 are negligible. Dykes up to 5 cm. in length.

40630 as 40629

40631 as 40629

40632 Shortclastic dykes in association with lenticular sedimentation units and disrupted laminae, T.

40633 as 40632

40634 Large clastic dyke with brecciated, angular mudstone inclusions.

40635 Clastic dykes showing extreme homogeneous shortening perpendicular to S_o $(+S_1)$ in the mudstone layers.

40636 Short clastic dykes associated with lenticular sedimentation units cross lamination. Silts contain carbonate, T.

40637 as 40632, T

40638 as 40632

40639 Extreme lenticular sedimentary units and clastic dykes with missing source beds.

40640 as 40629, T

40641 as 40629, T

The following are examples of dykes with compositional layers, and are from the Gordon Road section 47.5 miles from Maydena.

40642 Branching sub-hexagonal intersection pattern of clastic dykes and compositional layering.

40643 as 40642

40644 as 40642, T

40645 Branching pattern on one side and linear pattern on the other surface.

40646 Branching and linear form combined on one compositional surface.

40647 Very strong linear intersection pattern of clastic dykes and S_o.

40648 as 40647, very good development of fringe joint structures.

40649 as 40647

40650 as 40647

40651 as 40647

The following are $D_2 + D_3$ structures of the Gordon Road Section.

40652 Multiple clastic dykes quite strongly aligned in but generally slightly transected by S_3. Crenulation nature of the 3rd cleavage is often difficult to detect and has a slaty appearance in hand specimen.

40653 as 40652

40654 Single clastic dykes aligned in the S_3 direction or very close to it.

40655 as 40654, T

40656 as 40654, T x 2

40657 as 40654

40658 S_3 in varying lithologies ranging from mudstone to siltstone. In the more pelitic horizons it has a slaty rather than crenulation appearance, T

40659 as 40658

40660 as 40658

40661 as 40658

40662 as 40658

40663 as 40658

40664 as 40658

40665 as 40658

40666 S_3 at 25^o to S_o ($+S_1$) in fairly coarse grained pelite - still appears penetrative, T

40667 Crenulation nature of this cross cutting cleavage is clearly displayed, T, PT, PE

40668 as 40667

40669 Asymmetric F_2 in quartzite/chloritic phyilite assemblage with generated axial plane crenulation cleavage, T

40670 as 40669

40671 symmetric F_2 as above.

Specimens 40669 - 40671 are from 150m to East of the clastic dyke locality on the Gordon Road.

The following are from the Gordon River Road 47.5 miles from Maydena.

40672 Slump folds and clastic dykes in alternation of fine silts and mudstones.

40673 Slump fold and intraformational breccia in silt/mud.

40674 Intraformational breccia, angular mudstone fragments in siltstone.

40675 Disrupted and lenticular laminations.

40676 as 40675, including a possible slump fold.

40677 Load casts (?) and lenticular laminations accentuated by D_3.

40678 Parallel, fine laminations of mud/silt.

40679 as 40678

40680 as 40678, T

40681 Mud/silt alteration with irregular contacts (load casts), lenticular lamination and small clastic dykes, T

40682 Coarse silt layers with slump folds (?) and lenticular units.

40683 Mud/silt (with carbonate) alternation - S_3 in the thin mud layers and planar clastic dykes, T, CA.

40684 Lenticular sedimentary units probably scour and fill type related to penecontemporaneous erosion, T x 2

40685 as 40684

40686 as 40684

40687 as 40684

40688 as 40684

40689 Quartzite with garnets; these have statically overgrown S_1, some are very skeletal, T

40690 Phyllite with garnet pre-, and syn-D_2 and completely pre-S_3, T

40691 as 40690

40692 Amphibolite dyke found half way along the road section examined.

40693	Extreme examples of scour and fill structures.
40694	as 40693
40695	as 40693
40696	Clastic dykes and lenticular sedimentation units. Dyke/S_o intersection is linear.
40697	Large zone of completly slurried siltstone where any internal structure has been obliterated.
40698	Convolute lamination in a siltstone unit.
40699	Multiple liquefied zones in a finely laminated silt/mud.
40700	Examples of scour and fill structures. These penecontemporaneous erosion features were probably modified in shape during D_1.
40701	as 40700
40702	as 40700
40703	Coupled isocline of the strongly flattened IC type from the quartzite/chloritic phyllite sequence 150m East of the clastic dyke locality.

The following are from 1.8 kilometres along the Serpentine Dam Road, from the Gordon River Road.

40704	Strongly developed crenulation cleavage (S_3) in a dominantly phyllitic sequence, T
40705	as 40704, T
40706	as 40704
40707	as 40704, T x 2, PT.
40708	Clastic intrusives generally parallel to S_3 though most sedimentary structures have been obliterated, T
40709	as 40708, T, CA.
40710	as 40708 (4T)
40711	as 40708, T
40712	as 40708

Catalogue No.	Field No.	Grid Coords.	Preparations	Description
41705	7108056	16516654	T, PT	Phyllite
41706	7108071	16456670	T, PT,PE	Phyllite

The two previous specimens were used in a study of metamorphic conditions in the Gordon Dam - Strathgordon region which was reported in Boulter and Raheim (1974).

Specimens 43308 to 43312 are oosparite from Grunter Hill (55G DN 403002), Mayberry, Northern Tasmania. Strain analysis carried out on these specimens was reported in Chapter 7 and Boulter (1976).

Specimens 44366 to 44379 are quartz arenite from the eastern portion of the Frankland Range, Southwest Tasmania. Numbers underlined are those used in the field. Numbers in round brackets are field localities.

All of these specimens are quartz arenite and were measured to determine the 3D strain state. For each, several thin sections were cut denoted by A, B, C, 1st cut, XY, YZ, or XZ. The latter three are the principal sections of the finite strain ellipsoid, the first four characters are for sections cut at right angles to the XY plane with A and B usually perpendicular to one another and C their bisector. Full details of orientations can be obtained from Chapter 7.

44366	<u>7401025</u>	XY^2, XZ^2, YZ, 55G	DN33804154	(F87)
44367	<u>7401026</u>	A, B, C1, C2, XY	DN33804154	(F87)
44368	<u>7401028</u>	A, 1st cut, C, XY	DN34224097	(F92-F93)
44369	<u>7401029</u>	A, 1st cut, C, XY	DN34044130	(F94)
44370	<u>7401030</u>	A, B, C, XY Petrofabrics chip	DN29384076	(E86-85)
44371	<u>7401031</u>	A, 1st cut, B, XY	DN29354073	(E86-85)
44372	<u>7401032</u>	A, B, C, XY	DN30004006	(F24)

44373	7401036	A, B, C, XY	DN33574114	(F62)
44374	E86	XY, YZ, XZ	DN29384076	(E86)
44375	7401034	Petrofabric chip, XY, YZ, XZ	DN30004006	(F24)
44376	7302022	'XY', 'YZ', 'XZ'	DN25034360	(G94)
44377	7502001	XY, YZ, XZ	DN29214090	(P49)
44378	7502002	XY	DN29214090	(P49)
44379	7301055	XY,YZ, XZ	DN32784202	(F74)

The following specimens were used to characterise pre-strain shape factors.

44380	PEN 1	Dune cross bedded, millet seed sandstone with silica cement. Permian. COWRAIK QUARRY, PENRITH, ENGLAND. NY542310. Arrow point down dip of foresets.
44381	PEN 2	Layer immediately above Pen 1 (070.28)
44382	PEN 3	Two metres East of Pen 1/2 (070.28)
44383	PEN 5	Dune cross bedded sandstone. Quarry overlooking River Eden near the Courtfield Hotel, Appleby. Arrow is down dip of foresets.

44384	MEGS 1	LECKHAMPTON HILL QUARRY 50949186	1. (320.18) Oolitic limestone from a 1m thick cross bedded unit in upper level of the quartz.
44385	MEGS 2		2. Oolitic limestone (arrow on top bedding surface marks south) above Oolite Marl upper quarry level southern end.
44386	MEGS 3		3. (340.14) Oolitic limestone from within the very large cross bedded unit.
44387	MEGS 4		4. Well bedded oolite above Pea Grit in the Main Quarry.

44388	MEGS 5	CRICKLEY HILL 50932160	All oncosparites with varying amounts of ooliths. 5/6 are from the top of the Pea Grit, 7/8 80 cm below 5 and 6 within the Pea Grit.
44389	MEGS 6		7 T/S $\parallel S_o$ TWO T/S $\perp S_o$ (and mutually orthogonal)
44390	MEGS 7		
44391	MEGS 8		

44392 Bottle Weathered Pisolites

MEGS Specimens Ref: Geologists Association Guide No. 36, The Cotswold Hills, D.V. Ager. et. al., 1973.

From specimen 44381 to 44386 numbers of the type 070.28 are for orientation.

The following specimens are all from the Precambrian of the Frankland and Wilmot Ranges.

Catalogue No.	Field No.	Grid coords	Preparations	Description
46210	7105013	16706320	T	Quartzite
46211	7106007	20806280	T, TA	Amphibolite
46212	7106008	20806280	T, PT, PE	" "
46213	7106009	20806280	T, TA	" "
46214	7106018	16306870	T	" "
46215	7108011	16576600	T	Quartzite
46216	7108037	16786503	T	Phyllite
46217	72002	15926785	T, PT	"
46218	7202L	24004100	T	Meta-basite
46219	7201010	16376754	T	Phyllite
46220	7201011	18005013	T	Quartzite
46221	7201012	18005013	T	"
46222	7201013	18005013	T	"
46223	7201014	18105014	T	"
46224	7201018	17565001	T	"
46225	7201019	17474983		"
46226	7201025	17734864	T	Slight. mic. qtzite.
46227	7201026	17734864	T	" " "
46228	7201027	17734864	T	" " "
46229	7201033	18664802	T	" " "
46230	7201044	17004632	T	" " "
46231	7202025	22094527	T	Quartzite
46232	7202026	22094527	T	"
46233	7202027	21874519	T	"
46234	7202039	22924448	T	Slight. mic. qtzite.
46235	7202049	23184334	T	Phyllite
46236	7202050	23344318	T	Micaceous Qtzite.
46237	7202057	23984269	T x 2	Quartzose Phyll.
46238	7202061	20594318	T	Phyllite
46239	7204001	15926785	T	Qtzite./Phyllite
46240	7211025	19116423		Phyllite
46241	7211031	19486425	T	Gt. qtz. schist
46242	7211034	19526437	T	Qtz. phen. schist
46243	7211035	19526437	T	Meta-basite
46244	7301002	27324140	T	" "
46245	7301018	28414090	T	Quartzite
46246	7301019	28414090	T	"
46247	7301020	28624115	T	"
46248	7301022	28704088	T	"
46249	7301030	28484324		"
46250	7301037	29804090	T	"

Catalogue No.	Field No.	Grid coords	Preparations	Descriptions
46251	7301038	29624062	T	Quartzite
46252	7301041	29744005	T	"
46253	7301044	29454040	T	"
46254	7301045	30083913	T	"
46255	7301054	32984100	T	Slight.mic.Qtzite.
46256	7302004	32114032	T	Quartzite
46257	7302005	32124051	T	"
46258	7302006	34574122		"
46259	7302007	24284767		Micaceous Qtzite
46260	7302008	24074763	T	Slight.mic.Qtzite.
46261	7302009	22904728	T, PT	Quartzite
46262	7302014	24134683	T	Slight.mic.Qtzite.
46263	7302015	23804658	T	" " "
46264	7302017	23624840	T	" " "
46465	7302023	25004361	T	Quartzite
46266	7302024	20285586	T	"
46267	7302026	20285586	T	"
46268	7302030	20325595	T	Flaggy Qtzite.
46269	7302035	20465593	T	Schist
46270	7302041	20105338	T	Flaggy Qtzite.
46271	7302043	20175997	T, PT, PE	Qtz. Gt. Phen. Chl Chloritoid Schist
46272	7312004	15926625		Quartzite
46273	7401001	17025444	T	Flaggy Qtzite.
46274	7401002	16625430	T	" "
46275	7401003	16815300	T	Quartzite
46276	7401004	18385527	T	Flaggy Qtzite.
46277	7401006	19275423	T	" "
46278	7401007	19665404	T	" "
46279	7401012	17845916	T	" "
46280	7401013	17855935	T, PT, PE	Meta-basite
46281	7401014	20084592		Micaceous Qtzite
46282	7401015	19904948	T	" "
46283	7401016	20814992		Phyllite
46284	7401019	21165034		Qtzose. Phyll.
46285	7401020	20825034	T	" "
46286	7401021	20755022		" "
46287	7402002	17516011	T	Flaggy Qtzite.
46288	7402004	18785894	T	" "
46289	7402005	18855878	T, PT, PE	Micaceous Qtzite
46290	7402010B	16316199	T, PT, PE,CA.	Meta-basite
46291	7402012	15096169	T	" "
46292	7402013	15456153	T	Quartzite
46293	7402014	14596067	T	Phyllite
46294	7502003	28904095	T	Quartzite
46295	7502005	25004361	T	"
46296	7502009	24624338	T	"
46297	7502010	24054263	T	Phyllite
46298	7603002	14836650	T	Meta-basite
46299		14836650	T	" "

Catalogue No.	Field No.	Grid coords	Preparations	Description
46300	39166?	16706740	T, CA	Qtz. Carb. Phen. Chl Schist
46301	7108	16536612	CA	Phyllite
46302	7108038	16536612	CA	"
46303	7108045	16486622	T, CA	Qtzose. Phyllite
46304	7201001	16706320	CA	Micaceous Qtzite.
46305	7402008	19616097	T, CA	Garnet schist
46306	7506L	15926785	CA	Phyllite
46307	7302040	20005391	T, CA	Schist
46308	7108028	16826592	CA	Qtzose. Phyllite
46309		15926785		Quartzite-Phyllite

Preparation Key

T	-	Thin Section
PT	-	Polished Thin Section
PE	-	Probe Analysis
CA	-	XRF Analysis